

**Sediment flux and recent paleoclimate in Jordan Basin,
Gulf of Maine**

Lloyd D. Keigwin^{a*}

Cynthia H. Pilskaln^b

Submitted 3 Apr14 to Continental Shelf Research (CSR3289)
Resubmitted 28 July
Re-resubmitted 9 October

^aWoods Hole Oceanographic Institution
360 Woods Hole Rd.
Woods Hole, MA 02543

^bSchool for Marine Science and Technology
University of Massachusetts, Dartmouth

*corresponding author
telephone 508 289 2784
email: lkeigwin@whoi.edu

ABSTRACT

We report planktonic foraminiferal fluxes (accumulation rates) and oxygen isotopes ($\delta^{18}\text{O}$) from a nine-month sediment trap deployment, and $\delta^{18}\text{O}$ from three sediment cores in Jordan Basin, Gulf of Maine. The sediment trap was deployed at 150 m, about halfway to the basin floor, and samples were collected every three weeks between August 2010 and May 2011. The planktonic foraminiferal fauna in the trap is dominated by *Neogloboquadrina incompta* that reached a maximum flux in the second half of October. Oxygen isotope ratios on that species indicate that on average during the collecting period it lived in the surface mixed layer, when compared to predicted values based on data from a nearby hydrographic buoy from the same period. New large diameter piston cores from Jordan Basin are 25 and 28 m long. Marine hemipelagic sediments are 25 m thick, and the sharp contact with underlying red deglacial sediments is bracketed by two radiocarbon dates on bivalves that indicate ice-free conditions began 16,900 calibrated years ago. Radiocarbon dating of foraminifera indicates that the basin floor sediments (270-290 m) accumulated at >3 m/kyr during the Holocene, whereas rates were about one tenth that on the basin slope (230 m). In principle, Jordan Basin sediments have the potential to provide time series with interannual resolution. Our results indicate the Holocene is marked by $\sim 2^\circ\text{C}$ variability in SST, and the coldest events of the 20th century, during the mid 1960s and mid 1920s, appear to be recorded in the uppermost 50 cm of the seafloor.

Key words: Holocene, deglaciation, sediment trap, foraminifera, radiocarbon, oxygen isotope

Highlights:

Rates of sedimentation in Jordan Basin, Gulf of Maine, are sufficiently high (>300 cmkyr⁻¹) that interannual resolution is achievable in sediment cores.

The planktonic foraminifer *N. incompta* achieves maximum flux in late October, lives in the surface layer, and deposits its shell in oxygen isotope equilibrium with the sea surface.

N. incompta oxygen isotope composition can be used to reconstruct past sea surface temperatures (SST).

Within dating errors, late October SST was probably lower than today by 1-2°C during the 1920s and 1960s, and 2°C variability was typical during the past 5000 years.

1. INTRODUCTION

Basins on continental shelves are common where the margin was glaciated. In the northwest Atlantic Ocean, the Gulf of Maine is the southernmost occurrence of these basins (**Figure 1**) and it is also close to Canadian and New England centers of

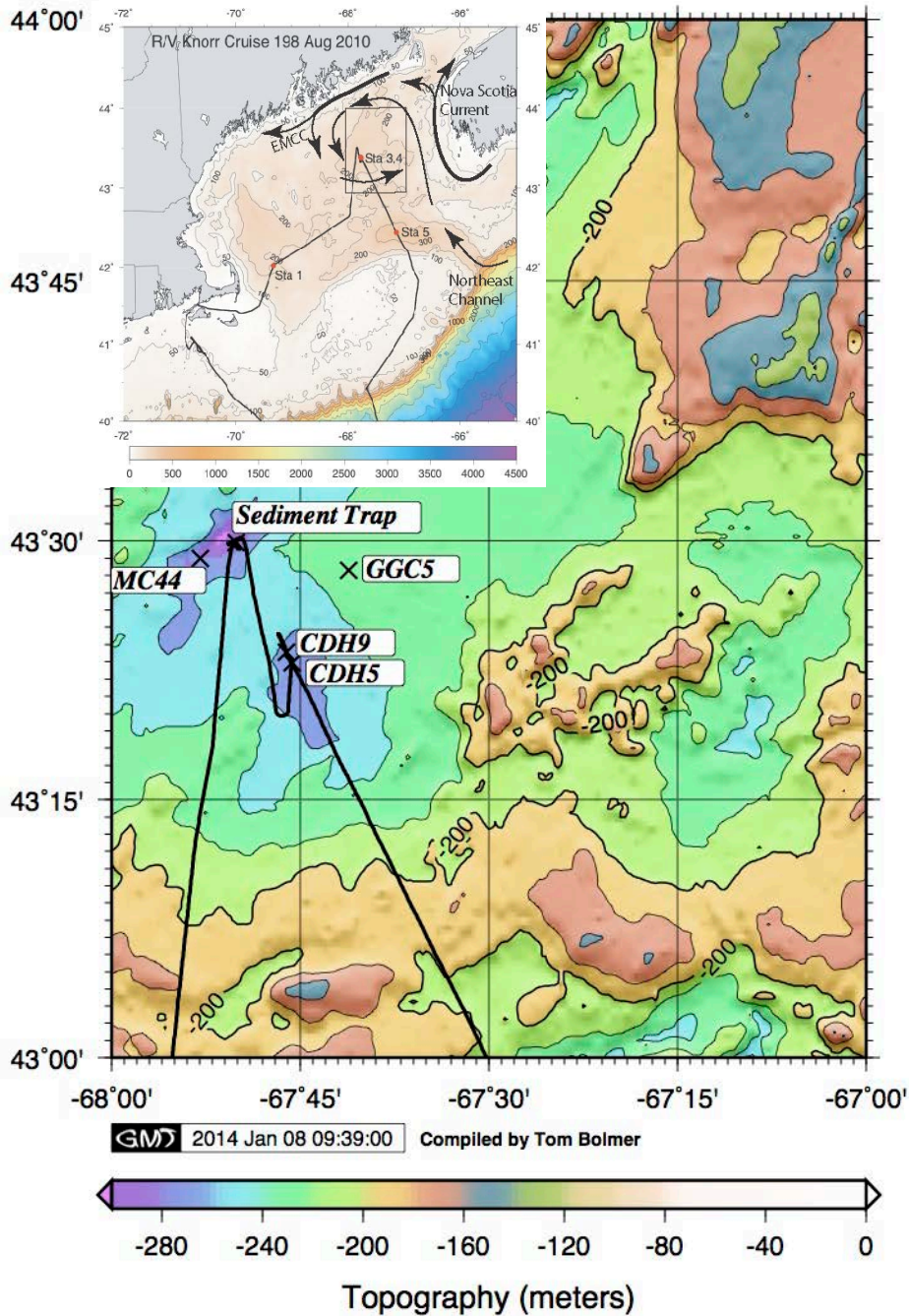


Figure 1. Locations discussed in this paper. Box in inset shows part of Jordan Basin in blow-up, with arrows illustrating surface currents relevant to Jordan Basin (the Eastern Maine Coastal Current, EMCC; the Nova Scotia Current; and flow through Northeast Channel). Additional cores in the study area are described by Schnitker et al. (2001).

oceanography. Because of its rich fisheries, investigations of the marine biology began in the 1870s by the U.S. Fisheries Commission and the world's first purpose-built oceanographic ship, *Albatross*, after 1881. Academic investigators began studying the

Gulf's circulation in the 1920s (Bigelow, 1927), and hydrographic data for Jordan Basin, in the northeastern Gulf, are available for the sea surface since 1870, and for deep water every year since 1965 (<http://www.bio.gc.ca/science/data-donnees/base/data-donnees/climate-climat-eng.php>; Brian Petrie, pers. comm. 2013). It was recognized long ago that major deep hydrologic changes result from intrusion of cold slope waters into shelf basins (Colton, 1968; Petrie and Drinkwater, 1993), forced by the North Atlantic Oscillation (NAO) (Petrie, 2007; Mountain, 2012). These data show substantially colder and fresher surface and deep waters in the mid 1960s than later years, and echo the changes observed at the same time in nearby Emerald Basin on the Scotian Shelf (Petrie and Drinkwater, 1993). Surface waters over Jordan Basin circulate cyclonically, and are cool and fresh, reflecting their Arctic source (Pettigrew et al. 2005) with additional contributions from local rivers and especially the Gulf of St. Lawrence (Houghton and Fairbanks, 2001). The strength of these surface flows, especially the Nova Scotia Current and the Eastern Maine Coastal Current, should affect upwelling in Jordan Basin as they define the Jordan Basin Gyre (Figure 1). Recent hydrographic studies have documented increased freshening in the coastal waters that flow southward past Nova Scotia and into the Gulf, and that such changes affect the entire ecosystem (Townsend et al. 2010). Over decades these are related to the NAO because when nutrient-rich warm slope waters flood the basins during the NAO positive phase, the productivity is higher (Drinkwater et al. 2003, Durbin et al., 2003). However recent freshening at high latitudes coupled with intense regional warming, has led to an overall decrease in phytoplankton and zooplankton abundance over the past decade despite the prevalence of positive NAO conditions (Pershing et al., 2005; Townsend et al. 2010; Mills et al., 2013).

Studies of particle flux from the surface ocean using sediment traps are a main source of information about the seasonal flux (mass accumulation rate) of particles, their transport, their chemical composition, and their role in chemical mass balance in the ocean. Sediment trapping is the only method to determine short-term accumulation rates of sediment. At most open ocean locations, the shells of pelagic micro-organisms such as foraminifera, coccolithophores, radiolaria and diatoms are an important part of the mass flux from the surface, and their study in sediment trap samples provides an important opportunity for paleoceanographers to ground truth or calibrate bio-geo-chemical proxies to surface ocean conditions. Surprisingly, there are few locations in the North Atlantic where both sediment trapping and sediment coring for long paleo records have occurred together. We are aware of only two examples: Cariaco Basin (Black et al., 1999) and the Sargasso Sea. At Sargasso Sea Station S (now known as the Bermuda Atlantic Time Series (BATS) (Deuser and Ross, 1989; Conte, 2001), trap data have been especially useful in calibrating proxy measurements such as Mg/Ca (Anand et al. 2003) and in enabling a late Holocene time series of SST (Keigwin, 1996).

Here we discuss a new sediment trap time series, and accompanying sediment cores, from Jordan Basin in the Gulf of Maine (Figure 1). The trap location was chosen specifically to complement sediment cores taken nearby, and to benefit from a nearby hydrographic station. The trap was about 2 km east of hydrographic buoy M that is a component of the Northeast Regional Association of Coastal and Ocean Observing Systems (NERACOOS; http://www.neracoos.org/realtime_map), formerly the Gulf of Maine Ocean Observing System (GoMOOS). Sediment coring in the Gulf of Maine has

occurred for decades, with the goal of revealing the history of glacial retreat (Tucholke and Hollister, 1973; for an overview, see Schnitker et al. 2001).

We focus on the recent climate history of Jordan Basin, based on changes in the stable isotope ratios in planktonic foraminifera. This would not have been possible without the synergy of both sediment trapping and sediment coring at the same location. We find that Jordan Basin is a remarkable archive of paleoclimate information, and that it should contain a history of climate phenomena such as the NAO that are not easily resolved at other locations off the continental shelf. Jordan and other shelf basins in the northwest Atlantic have generally not been exploited for their climate archives.

2. MATERIALS AND METHODS

Jordan Basin trap and core locations were occupied in August 2010 onboard cruise 198 of R/V *Knorr*. We approached the study area along a course surveyed previously by cruise 10-90 of R/V *Cape Hatteras* (Schnitker et al., 2001) and deployed a time-series sediment trap (model Mark 7, McLane Research Labs, Inc.; Honjo and Doherty, 1988) at 150 water depth (Figure 1). We then headed for two stations that are each within about a mile of a piston core studied by Schnitker et al., (2001), and recovered large diameter gravity cores (GGCs) and piston cores at each, and a multicore (MC) (**Table 1**). At 25.5 and 28.4 m in length, the piston cores are nearly double the length of the *Cape Hatteras* cores. The long cores are called “CDHs” in honor of the late Dr. Charles D. Hollister of Woods Hole Oceanographic Institution (WHOI) who pioneered long coring in the 1970s, including the Gulf of Maine (Tucholke and Hollister, 1973). The *Knorr* (KNR) cores were logged for magnetic susceptibility onboard the ship and the data are archived (<http://ncdc.noaa.gov/paleo/study/16974>). We also discuss results on cores OCE400 MC44 (here, MC44), and CH 12-04 GGC5 (here, GGC5) from the WHOI core collection. Unlike the other cores, GGC5 came from the slope of Jordan Basin, about 40 m above the basin floor.

Table 1. Locations discussed in this paper.

core	depth	core length (cm)	North latitude	West longitude
OCE400 MC44	287	53	43° 28.982'	67° 52.957'
CH1204 GGC5	231		43° 28.290'	67° 41.140'
KNR198 CDH5	277	2545	43° 22.935'	67° 45.671'
KNR198 CDH9	271	2842	43° 23.560'	67° 46.114'
sediment trap	290		43° 29.901'	67° 50.124'
Mooring M0117 (NOAA 44037)	285		43° 29.410'	67° 52.300'

2.1 Sediment trap

The time series sediment trap was deployed 7 August 2010 on a subsurface, bottom-anchored mooring in Jordan Basin, and it was recovered 2 May 2011. It has a baffled collection area of 0.5m² and samples were collected in thirteen 250 ml cups. Trap

cups were pre-poisoned with an 8% density-adjusted formalin solution in filtered seawater buffered to a pH of 8.0–8.1. Particulate sample handling and processing followed the standard procedures detailed in PilskaIn and Paduan (1992) and PilskaIn et al. (2004). Immediately following trap recovery, sample cup supernatant pH values were recorded, additional buffered formalin solution was added and all samples were stored at 4°C. Total cup samples were gently washed with filtered seawater through 1 mm Nylon mesh sieves to remove swimmers (e.g., copepods, euphausiids, amphipods, pteropods), and the remaining particulate sample material was quantitatively split into quarters using a four-section wet splitter (McLane Research Labs, Inc.). Two wet splits were filtered (Nucleopore polycarbonate, 0.4 µm), dried and weighed for mass flux determination and dried ground subsamples were analyzed for particulate inorganic carbon (PIC as CaCO₃) content using coulometry. PIC was determined by 1M phosphoric acidification of duplicate powdered samples in the coulometer, and measurement of evolved CO₂ (PilskaIn et al., 1996; 2004). The calibration standard used for the inorganic carbon analysis was 100% pure, dried CaCO₃; precision of the PIC analysis ranged from 0.3 to 1%. Planktonic foraminifera were picked wet, identified, and counted from the >150 µm sieve fraction of one ¼ split from each trap cup.

2.2 Sediment cores

Sediment cores were sampled differently, depending on our interests and resources at the time. MC44 was extruded, sliced into 1-cm slabs, and bagged. The bagged samples were then sub-sampled, dried, and weighed (mass range = 7-21 g). GGC5 was sampled at 4-cm spacing (2-16 g dry); CDH9 samples were smaller and taken initially at 50-cm spacing (5-10 g dry). Subsamples of all cores were washed in tap water over a 63 µm screen, dried, and sieved at >150 µm for picking and identification of planktonic foraminifera. Planktonic foraminifera were counted in every MC44 sample (data archived as above), and in GGC5 samples down to 120 cm, but not in CDH9 except for two samples that were dated. Some GGC5 samples needed to be split before counting.

In all cores, $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ were measured on 8-20 *Neogloboquadrina incompta* individuals from the size fraction 150-250 µm. Stable isotope measurements were made on a VG Prism instrument at the National Ocean Sciences Accelerator Mass Spectrometry (NOSAMS) facility at WHOI. Instrument precision is about 0.05 ‰ for both isotope pairs, but the precision on foraminiferal samples is nominally about 0.10 ‰. NOSAMS also did the radiocarbon dating on planktonic foraminifera and bivalves from these cores (1.4 to 7.0 mg). An age model was developed for each core after conversion to calibrated ages using CALIB 6.1 (Stuiver and Reimer, 2011) with no additional reservoir correction. Ages between dated samples were linearly interpolated.

3. RESULTS

3.1 Total mass and CaCO₃ fluxes in sediment trap

The time series of total particulate mass flux (TMF) at 150 m displayed primary peaks in August and October, with a secondary peak in November. Fluxes steadily increase in the spring (**Figure 2a**). Measured carbonate fluxes revealed a peak in August

followed by a large flux maximum in the late fall (November)(Figure 2b). The August CaCO_3 flux represented 31% of the total mass flux (TMF); the November carbonate flux contributed 76% of the measured TMF for that month.

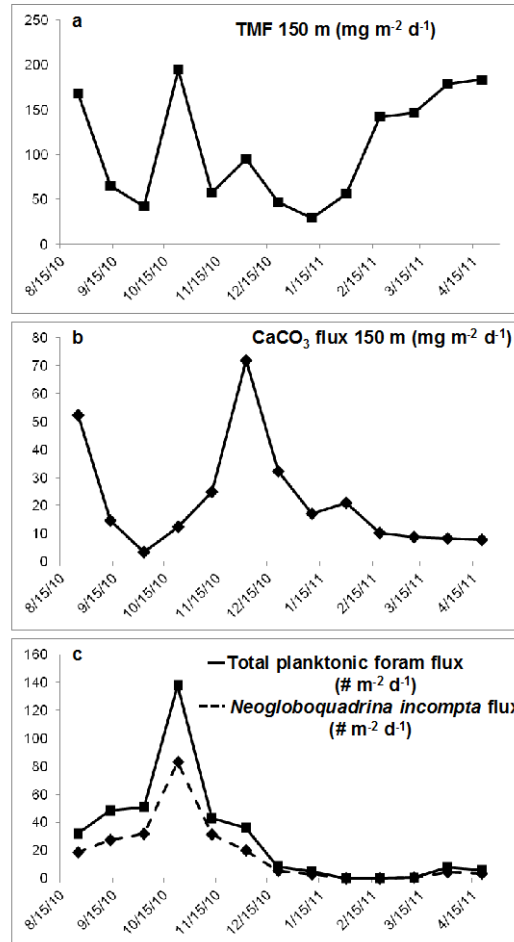


Figure 2. (a) Total Mass Flux (TMF), (b) carbonate flux, and (c) foraminiferal fluxes in Jordan Basin, 2010-2011.

Foraminiferal fluxes differ from the carbonate flux (Figure 2c). In August the foraminiferal flux is low, and the peak foraminiferal flux occurred in October, about a month before peak carbonate flux. The August CaCO_3 peak at the beginning of the time-series may be due to the abundance (not quantified) of small, larval bivalves in the sediment trap cup. It is also likely that coccoliths could have contributed carbonate material to the August flux. Summer blooms of *Emiliana huxleyi* are observed in the gulf (Townsend et al., 1994), with trap and surface sediment coccolith accumulation documented by alkenone biomarker studies (Prah et al., 2001). Our light microscope examination of the trap samples collected in the present project does not allow for the identification or counting of the micron-sized coccoliths.

3.2 Planktonic community observations

3.2.1 Seasonality in trap samples.

Our 8-month series of trapped planktonic foraminifera is dominated by *N. incompta* that accumulated between August 2010 and March 2011 (Table 2). Peak abundance of this species occurred during the second half of October, when the number was nearly 3 times that of August to September and November to December. After *N. incompta*, the next dominant taxon is *Globigerina bulloides* and those were more evenly distributed between August and November with a small peak in October. The $\delta^{18}\text{O}$ of trapped *N. incompta* ranged from -0.62 (early October) to 0.25 ‰ (early December)(Table 3).

Table 2. Sediment trap fauna and flux (accumulation rate; (number)/m²/day) of *N. incompta*

mid point date	No. days	<i>Globorotalia inflata</i>	<i>Globigerina bulloides</i>	<i>N. incompta</i>	<i>Neogloboquadrina</i> left coiling*	<i>Turborotalia quinqueloba</i>	<i>Globigerinita</i> sp.	<i>Globigerinita glutinata</i>	unident	total PF**	flux <i>N. incompta</i>	total PF flux
25-Aug-10	19	1	11	31	1	0	0	0	0	44	13	19
13-Sep-10	20	2	12	53	0	0	1	0	0	68	21	27
3-Oct-10	20	0	30	48	1	0	0	0	0	79	19	32
23-Oct-10	20	12	54	137	1	3	1	0	0	208	55	83
12-Nov-10	20	13	33	29	1	0	0	1	1	78	12	31
2-Dec-10	20	3	6	40	0	0	0	1	0	50	16	20
21-Dec-10	19	1	3	8	0	0	0	0	0	12	3	5
10-Jan-11	20	0	2	5	0	0	0	0	0	7	2	3
30-Jan-11	20	0	0	0	0	0	0	0	0	0	0	0
19-Feb-11	20	0	0	0	0	0	0	0	0	0	0	0
11-Mar-11	20	1	0	0	0	0	0	0	0	1	0	0
31-Mar-11	20	1	2	8	0	0	0	0	0	11	3	4
20-Apr-11	19	0	0	6	2 N.p.	0	0	0	0	8	3	3

*note: of left-coiling *Neogloboquadrina*, only two (April) were typical *N. pachyderma*

**PF= planktonic foraminifera

Table 3. Stable isotope results and flux-weighted mean $\delta^{18}\text{O}$ of *N. incompta*

collection mid point	$\delta^{13}\text{C}$ ‰	$\delta^{18}\text{O}$ ‰	<i>N. incompta</i> #/m ² /d	flux-weighted $\delta^{18}\text{O}$
25-Aug-10	0.36	-0.44	13	-0.039
13-Sep-10	0.45	-0.59	21	-0.087
3-Oct-10	0.44	-0.62	19	-0.083
23-Oct-10	0.44	-0.42	55	-0.159
12-Nov-10	0.47	0.13	12	0.011
2-Dec-10	0.36	0.25	16	0.028
21-Dec-10	0.28	0.11	3	0.003
10-Jan-11			2	
30-Jan-11			0	
19-Feb-11			0	
11-Mar-11			0	
31-Mar-11	0.29	0.17	3	0.002
20-Apr-11			3	
Total flux= 142				
mean of flux-weighted $\delta^{18}\text{O}$ =				-0.041

Lagging the abundance maximum in planktonic foraminifera by about a month, a pulse of the coiled pteropod *Limacina retroversa* registered more than a thousand in the sample centered on 2 December 2010. This is the only large sample of pteropod that was split and counted (<1mm). This event extended through January 2011, and is coincident with the high flux of CaCO_3 that extended from November through January. *L.*

retroversa is commonly observed in Gulf of Maine plankton tows and displays increased abundance in the late fall-winter months (Bigelow, 1926; Redfield, 1939; Durbin et al., 2003).

3.2.2 Faunal change down-core.

Sediment core results are consistent with sediment trap results in that the seafloor record is dominated by the abundance of *N. incompta*. In MC44, the abundance of planktonic foraminifera is relatively high (tens per g) near the top and the bottom of the core, but between those peaks it is low ($<10\text{g}^{-1}$) (**Figure 3**). In MC44 *N. incompta* are typically about 90% of the fauna, excepting an interval around 15 cm where abundance drops to about 65%. Those samples have more *G. bulloides*, *T. quinqueloba*, and *N. pachyderma* (data archived as above). The latter species rarely exceeds 10% in MC44. The fauna of GGC5 differs significantly from that of MC44: (1) the relative abundance of *G. incompta* only approaches 90% at the core top and declines below that, and (2) the % *N. pachyderma* follows the opposite trend with a doubling of abundance below about 50 cm. Although the foraminifera at CDH9 were not counted, *N. incompta* is continually present above 1900 cm and there are only rare occurrences below that.

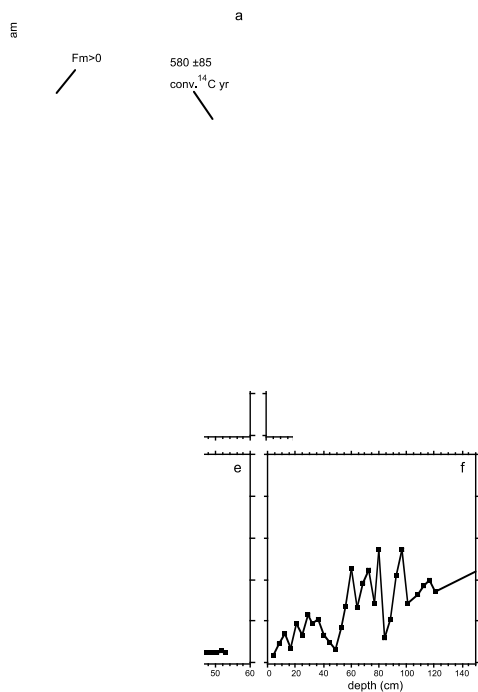


Figure 3. Planktonic foraminiferal counts from the multicore at the bottom of Jordan Basin (left column), and the gravity core on the basin slope (right column). AMS ^{14}C dated levels are shown in panels a and b.

Fm>1 indicates that the sample contains bomb radiocarbon. Based on age-depth relationships (Figure 3), there is about a 1000 year gap between the bottom of MC44 and the top of GGC5. The gravity corer probably blew by uppermost sediments, whereas the multicorer was designed to recover a perfect sediment-water interface.

3.3 Radiocarbon dates and sedimentation rates in Jordan Basin

Sedimentation rates are mostly based on AMS ^{14}C dates for Jordan Basin cores, hence their age models are only approximate because there are so few dates (**Table 4**; Figure 4). However, despite this uncertainty, the indicated rates are so high that they are notable. MC44 is the best dated of the cores. The top is zero age (Fraction modern>1),

which is a clear sign of the presence of excess ^{14}C produced by atmospheric testing of thermonuclear weapons that peaked in the 1960s. At the bottom of MC44 the conventional ^{14}C age is 580 ± 80 years. These are the only two levels where the abundance of planktonic foraminifera is sufficient for the AMS measurement (Figure 3). When the bottom age of MC44 is calibrated, assuming $\Delta R=0$, the result is 198 ± 102 years B.P. That result would give a sedimentation rate of ~ 230 cm/kyr, except that independent evidence suggests it must be higher. Dr. E.A. Boyle (MIT) has sampled MC44 for studies of Pb isotopes (pers. comm. to LDK, 2013). ^{210}Pb dating down to 32 cm indicates an age of roughly 55 years (580 cm/kyr), but it does not constrain the accumulation rate below that depth. $^{206/207}\text{Pb}$ is more interesting. Boyle hypothesized that Jordan Basin is a site that should have the atmospheric signal of mining and smelting that began in the upper Mississippi Valley (UMV) between about 1840 and 1850. That airborne signal is found at highly resolved sites downwind, for example, the Pettaquamscutt River in Rhode Island (Lima et al., 2005). Near the bottom of MC44, $^{206/207}\text{Pb}$ data indicates a peak that may be the UMV emission maximum. Thus, we assume the bottom of MC44 is only about 150 years old, which is within the errors of the AMS ^{14}C measurement. That gives a sedimentation rate of about 330 cm/kyr for the upper half meter of the Jordan Basin floor.

Table 4. AMS radiocarbon results on cores from Jordan Basin, Gulf of Maine

Core Information	Sample Description	Mass mg	NOSAMS	Age Conv. ^{14}C yrs	Age error years	Calib 6.1 Age, yrs.	Calib error years
			Accession Number				
OCE400 MC-44							
3.5 cm	mixed planktonic sp.	3.6	OS-66198	>Mod		0	
50 cm	mixed planktonic sp.	1.4	OS-66977	580	85	198	102
CH1204 GGC-5							
44.5 cm	mixed planktonic sp.	2.3	OS-76270	2480	40	2132	74
276.5 cm	<i>N.pachyderma</i>	1.4	OS-74857	9750	70	10606	78
KNR198 CDH-9							
100 cm	mixed planktonic sp.	1.9	OS-103200	1630	20	1201	62
1550 cm	mixed planktonic sp.	2.8	OS-103198	5070	25	5427*	43
2550 cm	bivalve (cf <i>Abra</i>)	7	OS-102837	13850	65	16638	141
2580 cm	bivalve (cf <i>Abra</i>)	2.4	OS-103199	14450	45	17064	123

*15% chance 5359 ± 12 ; 0.6% chance 5335 ± 1

Based on AMS dates in CDH9, for the past 5 kyr the linear rate of sedimentation on the basin floor was ~ 340 cm/kyr (Figure 4). Beyond that the apparent rate decreases to ~ 90 cm/kyr, but without further dating the inflection point is uncertain and the age model must be inaccurate. However, we base no conclusions on this poorly constrained part of the basin floor section. The two oldest dates in CDH9 constrain the distinct lithological change from dark gray-brown hemipelagic sediment above to brick-red sediment below.

Basin slope core GGC5 has a ten-fold lower sedimentation rate than the basin floor cores (Figure 4). Both the slope and basin floor sites are overlain by thick, highly turbid benthic nepheloid layers (BNLs). The deep eastern gulf/Jordan Basin BNLs are

thought to be maintained by bottom sediment resuspension (via tidal mixing) and possible downslope transport of fine sediment from the seafloor above about 100 m water depth (Pilskaln et al. 2013a,b). Thus, the fact that GGC5 has a much lower rate of sedimentation than the basin floor cores probably indicates that sediment from the BNL preferentially settles out on the basin floor, or that sediment deposited on the slope can be mobilized by a process such as bioturbation.

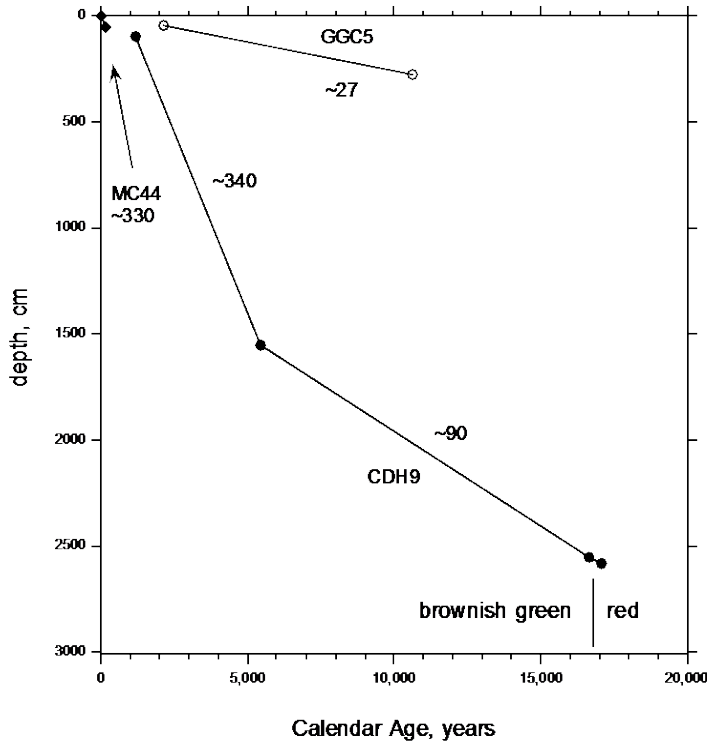


Figure 4. Rates of sedimentation (cm/kyr) in Jordan Basin sites. Note that GGC5 has a much lower rate than the other sites because it is on the basin slope and thus is a source of sediment to the basin floor.

3.4 Oxygen isotope stratigraphy

The $\delta^{18}\text{O}$ record from MC44 reveals no systematic change over its length (**Figure 5**). Its most distinctive feature is the generally increased values between about 15 and 35 cm that define an excursion of about 0.5 ‰. On average, the upper 80 cm of GGC5 has $\delta^{18}\text{O}$ of about 0.4 ‰, similar to results at MC44 but without anything similar to the 0.5 ‰ excursion. This suggests it bypassed the youngest sediments, as often happens with gravity cores and piston cores. Below 85 cm, GGC5 $\delta^{18}\text{O}$ decreases to minimum values of -0.4 ‰ with variability of about 0.4 ‰. The much longer and less closely sampled record of CDH9 also becomes isotopically lighter down core, but not to the extent of GGC5. The amplitude of $\delta^{18}\text{O}$ events in CDH9 exceeds 0.5 ‰, making it more like MC44 than GGC5.

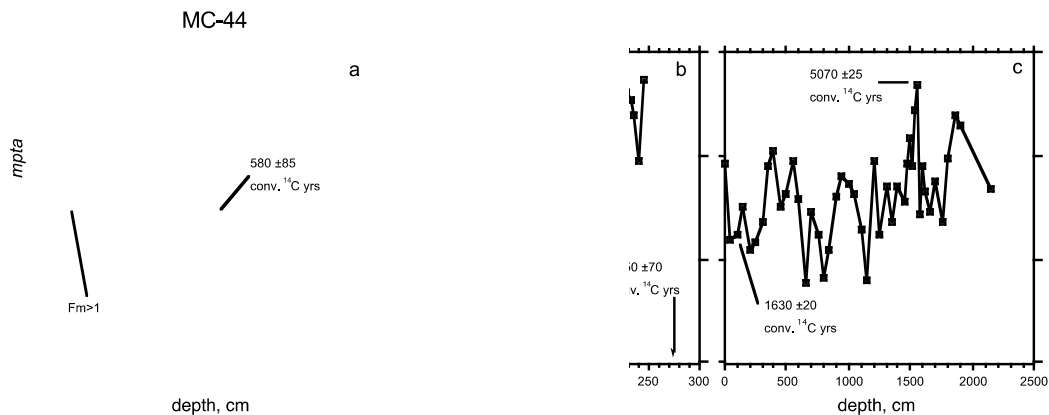


Figure 5. Oxygen isotope ratios on *N. incompta* in the three cores of this study. Note the different depth scales, and that maximum $\delta^{18}\text{O}$ is only reached in MC-44.

4. DISCUSSION

4.1 Variation in TMF, CaCO_3 and planktonic foraminifera fluxes

Time-series sediment trap data at 150 m in Jordan Basin have been collected previously using the same sediment trap design and deployed at essentially the same location as in the current study. Trap flux data from Sept. 1995 to May 1997 (obtained as a component of the NOAA Gulf of Maine Regional Research Program) showed CaCO_3 peaks in Sept.-Oct. 1995, Feb.-March 1996 and Oct.-Nov. 1997; all were coincident with increases in the TMF (Pilskaln, 2009; Pilskaln et al., 2013). Because those trap deployments showed insignificant carbonate flux during the late spring and summer, we decided to not trap from May to August in order to achieve higher temporal resolution during the remaining months. Approximately 10 years later (April 2005 to April 2006), a 150 m trap (same design as used in herein) was deployed as part of the Gulf of Maine Ecology and Oceanography of Harmful Algal Blooms (ECOHAB) Program. The 12 month time-series revealed a CaCO_3 flux maximum in Aug. 2005 and another peak in Feb.-March 2006, with the TMF displaying peaks at the same time, although the maximum TMF was observed in April-May 2005 (Pilskaln et al., 2013a). Unfortunately, planktonic foraminifera were not counted in these earlier trap programs, and they have dissolved in archived samples. However, microscope observations noted abundant planktonic foraminifera in the fall trap cups that were very likely *N. incompta* based on a few photographs of the samples.

Clearly there is significant interannual variability in the seasonal timing of the carbonate flux peaks in the eastern gulf provided by foraminifera, pteropods and/or coccoliths as demonstrated by three trap studies at the same location and depth over a period of ~15 years. Additionally, the maximum CaCO_3 fluxes in an individual year are not necessarily coincident with TMF peaks. But of greater interest here from a sedimentology and micropaleontology viewpoint is the fact that the planktonic foraminifera are the only carbonate flux components to be preserved in the sediment

record. Although collected abundantly in gulf traps deployed above the pervasive BNLs, aragonitic pteropod shells are consistently absent from the particulate material obtained in the BNLs, and are not found in the underlying sediments (Pilskałn, 2009; Hayashi and Pilskałn, 2013). The reasonable conclusion from these observations is that the sinking shells dissolve at depth in the gulf because aragonite is more soluble than the calcite foraminifera and coccoliths. This is supported by profiles and time-series measurements showing low omega aragonite (<1.3), low pH (<7.9), and low buffer capacity (TA:DIC <1.1) in near-bottom waters (Wang et al., 2013; Salisbury, unpubl.). Coccoliths are also absent from gulf sediment cores, however the documentation of specific haptophyte alkenone biomarkers in gulf sediments indicates that the coccoliths (specifically from *E. huxleyi*) are deposited but largely dissolve as a result of early diagenesis (Christensen, 1989; Prahl et al., 2001).

4.2 Deglacial and Holocene climate change

Core CDH9 recovered a deglacial lithological sequence similar to those described by Schnitker et al. (2001). This is not surprising as those authors show upper reflectors in deglacial units of Jordan Basin to be parallel and continuous. Holocene sediments are acoustically transparent, consistent with enhanced deposition of clays and silt with minor influence of ice. All cores have a long sequence of low magnetic susceptibility (mostly Holocene) that overlies a unit of much higher susceptibility. The main difference is that in CDH9 the low susceptibility interval is 25 m thick, about twice that in other cores discussed previously (Schnitker et al. 2001). At both CDH9 and other cores the major lithological change from upper gray-brown sediment to underlying red sediment is coincident with a ten-fold increase in magnetic susceptibility. Schnitker et al. (2001) call this transition “Distal Glacial Marine (DGM)” to “Marine”, but they do not comment on the distinct red color in what must be DGM. Their ^{14}C date on the magnetic susceptibility change is 13425 ± 180 conventional ^{14}C yrs, based upon mixed benthic foraminifera. This result is ~ 400 years younger than our bivalve date on the oldest occurrence of the Marine facies (Table 3). Together with the bivalve date in the youngest red sediments, we constrain this lithological change to have occurred 14150 ± 300 conv. ^{14}C years ago. Schnitker et al. (2001) did not calibrate their ^{14}C date but instead adjusted it using a reservoir age of 600 years. They reasoned that coastal waters would be upwelling and older, but we have calibrated our dates with no ΔR because the circulation would have been strongly estuarine and the large flux of glacial meltwater probably would have prevented upwelling. If that is correct, our best estimate for the sudden disintegration of the ice shelf over Jordan Basin and the onset of marine sedimentation, as described by Schnitker et al. (2001), would have been ~ 16.9 calibrated kyr ago. This date is compatible with results from Owen Basin in southern Bay of Fundy that show the abandonment of the Owen Basin Moraine was several centuries earlier (Todd and Shaw, 2012). It makes sense that the onset of marine hemipelagic sedimentation in Jordan Basin came shortly after the onset of what is equivalent to DGM sedimentation in the Bay of Fundy.

Aside from the change from DGM to marine sedimentation, our data do not reveal much else about the early deglacial history of Jordan Basin because our sampling interval is coarse and the abundance of foraminifera limits $\delta^{18}\text{O}$ measurement. It is clear that late deglacial and maybe early Holocene sediments accumulated at a lower rate than most of

the Holocene (Figure 4). *N. incompta* is continually present to 19.0 m down the core, a level that is about 9.5 ka based on linear interpolation between the dated levels. Older than that it appears only once more in sufficient abundance for $\delta^{18}\text{O}$. Holocene $\delta^{18}\text{O}$ displays a long term increase that may signify gradually increasing salinity due to reduced ice melt, although there must be strong, unresolved signals that have been aliased by the wide sample spacing (Figure 6).

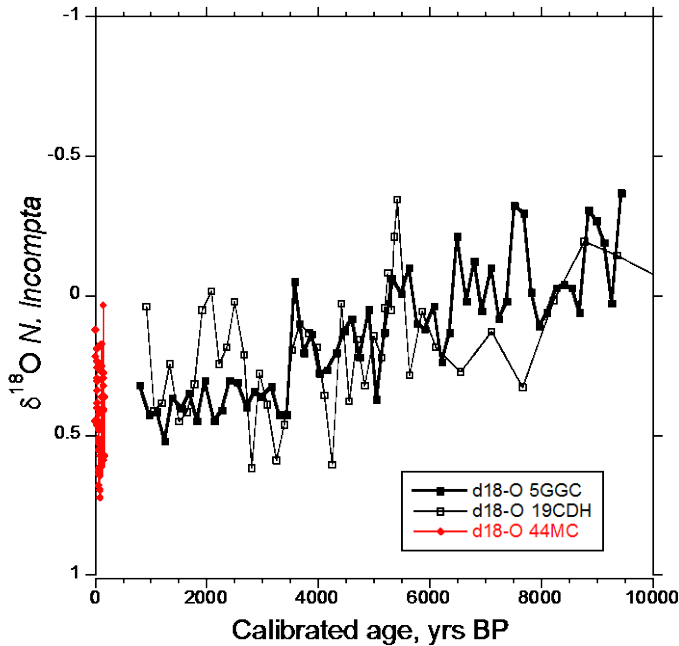


Figure 6. Composite of $\delta^{18}\text{O}$ records of three cores from Jordan Basin on a common age scale based in results in Figure 3. If the decade to centennial scale variability of $\sim 0.5\text{‰}$ in MC44 is typical of the Jordan Basin setting (see Fig. 8 for an expanded view), then with the coarse sampling CDH9, the lower resolution of GGC5, and the sparse ^{14}C dating, it is not surprising that the two longer data series have little in common.

High-resolution sampling in Jordan Basin cores and those from other basins on the continental shelf have great potential to reveal new details of climate change. For example, the prominent freshening that is expected around 8.2 ka (Condrón and Winsor, 2011) is not resolved in the $\delta^{18}\text{O}$ data, although we sampled closely around the event dated to 5.4 ka because we thought that might be it. If the 8.2 ka event lasted 160 years, as reported by Thomas et al. (2007), then given the assumed linear sedimentation rate in Jordan Basin it should occur around 18 m down core, within a maximum thickness of ~ 14 cm. Of course, the early Holocene rate of sedimentation may have been higher so the event may be found deeper in the core with a greater thickness. If $\delta^{18}\text{O}$ at MC44 is any guide, Holocene events of $\sim 0.5\text{‰}$ amplitude are the rule. The longer and higher sedimentation rate record of CDH9 may have captured some of these better than GGC5 (Figure 6).

4.3 Seasonality and equilibrium precipitation of calcite by *N. incompta*

Our sediment trap results provide essential information about the seasonality of planktonic foraminifera and where they live in the water column. This information can be gleaned only by sediment trapping or plankton towing, and when available it can lead to estimation of sea surface paleotemperatures (Section 4.4). As noted in section 3.2.1, the peak abundance of *N. incompta* in Jordan Basin occurs in October. Raw counts of

individuals are converted to accumulation rates by dividing the total number of individuals per cup by the surface area of the trap and by the number of days of collection. During the October abundance peak the flux is about 55 shells $\text{m}^{-2}\text{d}^{-1}$ (Figure 7, Table 2). Knowing this is the first step in determining where in the water column this species lives. The process we follow is similar to that described by Jonkers et al. (2010) for a sediment trapping location in the Irminger Sea.

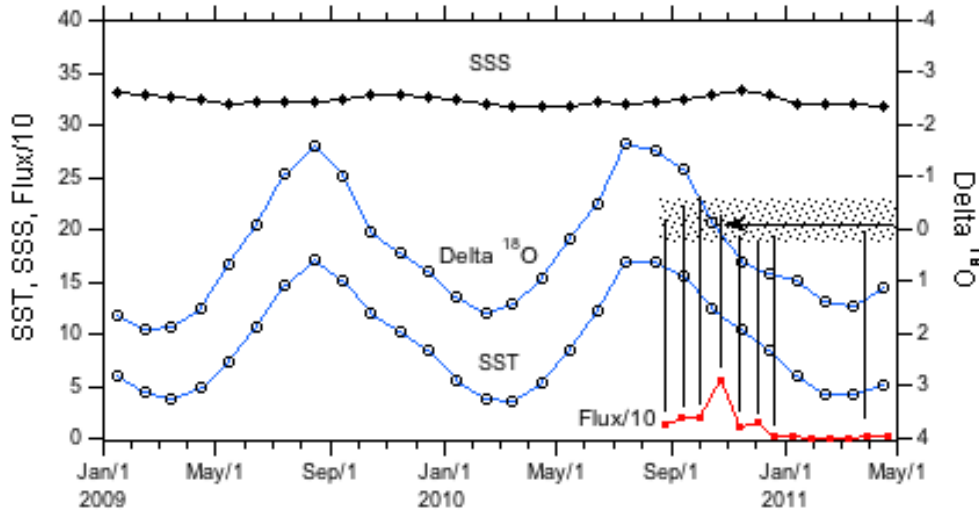


Figure 7. The relationship between hydrographic properties in Jordan Basin (SSS, SST), the $\delta^{18}\text{O}$ of calcite precipitated in equilibrium with surface water, and the flux (accumulation rate) and $\delta^{18}\text{O}$ of *N. incompta*. As described in section 4.3, $\delta^{18}\text{O}$ of calcite precipitated in equilibrium with SST and SSS is calculated from a paleotemperature equation, and that result can be compared to the measured $\delta^{18}\text{O}$ in foraminifera caught by a sediment trap. In this case, the flux of *N. incompta* plotted in red (and divided by 10 along the left axis) was maximum in late October 2010. The range of $\delta^{18}\text{O}$ of trap *N. pachyderma* d. is shown by the stippled pattern, and the results from individual trap cups can be read by how far up the y-axis the black lines extend from each flux point. The flux-weighted mean of the $\delta^{18}\text{O}$ observations (arrow at -0.04 ‰) is insignificantly different from the predicted equilibrium $\delta^{18}\text{O}$ for late October of 2010.

The second step is to determine the $\delta^{18}\text{O}$ of surface seawater. This has been calculated using the observed monthly average sea surface salinity and sea surface temperature (SST) from the NERACOOS Buoy M in Jordan Basin (2010-2011), and the basic North Atlantic $\delta^{18}\text{O}$ -S relationship ($d\delta^{18}\text{O}/dS=0.60$) of Craig and Gordon (1965) (Figure 7). This is only an approximation. There are many sources of fresh water and seawater in the Gulf of Maine, and they all have different equations with different slopes (Houghton and Fairbanks, 2001). For the Northeast Channel, the conduit for most water to the Gulf of Maine, $d\delta^{18}\text{O}/dS=0.57$ for the time between June 1995 and July 1997 (Houghton and Fairbanks, 2001). This relationship would give similar results to our assumed relationship but as we do not know *a priori* how and when it might have changed in the past the conservative assumption is to use a North Atlantic average. With this information, the third step is to predict the $\delta^{18}\text{O}$ of calcite precipitated in equilibrium with surface seawater using an equation such as that developed by Shackleton (1974). Using that equation results in the $\delta^{18}\text{O}$ curve shown in Figure 7.

The fourth step is to estimate the mean value of the $\delta^{18}\text{O}$ of *N. incompta* during the seasons it grows. To do this we weight the $\delta^{18}\text{O}$ measurements according to the flux of *N. incompta* in samples where there were enough to measure. We find that the flux-weighted mean $\delta^{18}\text{O}$ of *N. incompta* (-0.04 ‰) is close to the calculated equilibrium value for October (0.06 ‰) when the flux is maximum (Figure 7). This value has been adjusted to account for the difference in timing of the hydrographic observations (15 October) and the flux maximum (23 October). The difference between the two (0.10 ‰) is close to the precision of the analysis. Thus, we conclude that on an annual basis *N. incompta* grew its shell close to equilibrium with Jordan Basin surface seawater with the flux centered on the last half of October 2010.

Here we must introduce a caveat. Whereas we conclude this species can be thought of as living, on average, in surface seawater, it does not necessarily live at the sea surface. This is because deep mixing begins in October in Jordan Basin and waters of about 12°C first outcrop at the surface then (Brown and Irish, 1993). Thus, from the data of Brown and Irish (1993), it could have lived as deep as 75 m in October and even deeper later in the year. This contrasts with results from other North Atlantic studies. For example, in the Irminger Sea, maximum foraminiferal flux occurs during summer stratification (Jonkers et al., 2010), whereas in the Gulf of Maine the maximum occurs when stratification breaks down. In nearby waters it was reported that *N. incompta* grew close to the sea surface south of Cape Cod during the winter (Keigwin et al. 2005), and studies off Ireland show it dominates the fauna in spring and summer (Chapman, 2010; Harbers et al. 2010).

4.4 Estimation of SST from $\delta^{18}\text{O}$ of *N. incompta*

At this point we know the average $\delta^{18}\text{O}$ of *N. incompta* reflects the surface hydrographic conditions in Jordan Basin in late October, but in order to derive SST from the foraminiferal $\delta^{18}\text{O}$ we need to know the relative roles of SST and SSS in the $\delta^{18}\text{O}$ of equilibrium calcite. A simple regression analysis of the data shows that $\delta^{18}\text{O}$ calcite is a significant linear function of SST ($r^2=0.95$) but not of SSS ($r^2=0.00$). At first this seems surprising, but applying the simple rule of thumb that a 1 ‰ change in $\delta^{18}\text{O}$ corresponds to a ~4°C change in temperature (Figure 7) accounts for this very strong relationship (also see discussion of Jonkers et al. 2010).

We derive a history of SST only for MC44 because that record is closest to modern where our proxy calibration is most likely to be accurate. The longer histories of GGC5 and CDH9 have the ice volume effect embedded as well as the effect of local flux of fresh water from melting of nearby ice masses. For the MC44 data, we first apply the empirical relationship $\delta^{18}\text{O} = -.24\text{SST} + 2.59$ from the above analysis. However, the resulting SSTs require an important adjustment. Note that $\delta^{18}\text{O}$ in MC44 ranges from ~0.0 to 0.7 ‰ (Figure 5), whereas predicted equilibrium values for today's conditions are lower by about 0.5 ‰ (stippled pattern in Figure 7), and the two ranges overlap by no more than 0.25 ‰. This is typical of seafloor samples because the more fragile and isotopically "lighter" members of a population often are fragmented or dissolve and disappear altogether from the geological record (Berger and Killingley, 1977).

Using a 2°C adjustment to calculated SSTs (from the "rule of thumb" for a 0.5‰ change) and a linear age model based on a zero age core top and 150 yrs at 50 cm

(section 3.3), the derived SST record from MC44 is compared to 1° by 1° gridded instrumental observations for October over the past 150 years (**Figure 8**). The coldest epochs of the past century (~1925 and ~1965) seem to be recorded by Jordan Basin sediments, and the more recent of them corresponds to 1960s minimum in the NAO (Hurrell, 1995) and the Great Salinity Anomaly that began near the end of that decade (Dickson et al. 1988). This age model may make the cool events 10 or 20 years older than the instrumental record, but the important points are (1) that Jordan Basin sediments have decadal scale resolving power even if we may not be able to date events to within a decade, and (2) that our paleo-SST calculations give reasonably reliable results.

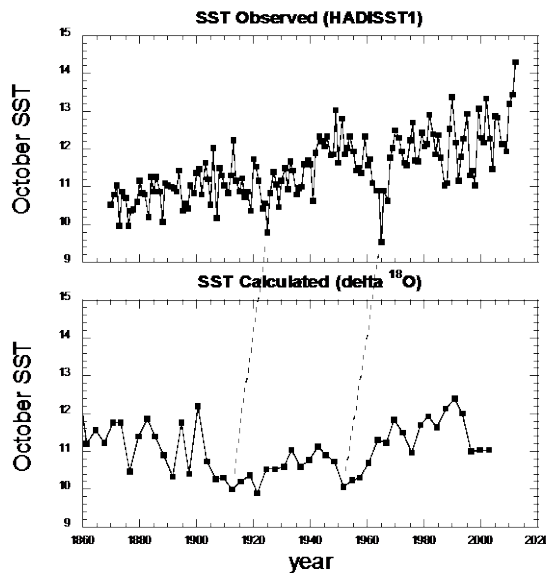


Figure 8. October sea surface temperature (SST) based on methods described in section 4.4 (lower panel), compared to gridded instrumental observations since 1870 (upper panel, data from HADISST). Dashed lines suggest a correlation between the foraminifera based isotope method and instrumental data. If this is correct, it would indicate the age model is too old by 10 to 20 years, a difference that is probably well within the error of our methods.

Unfortunately there are few other climate records from the Gulf of Maine with which we can compare our results. The closest and most detailed climate record comes from near the Maine coast, west of Jordan Basin (Wanamaker et al., 2007). That result is based on $\delta^{18}\text{O}$ in three bivalves. Each records a temperature series of about century duration, and together they form a discontinuous history of 2°C cooling over the past 1000 years. Wanamaker et al. (2007) interpret that history in terms of either cooling of the Labrador Current during the millennium, increased transport of that current, or reduced influence of warm slope waters. Importantly, they link these phenomena to the role of the North Atlantic Oscillation in the Gulf of Maine, and this conclusion is consistent with other studies that take a much broader view of the western North Atlantic (for example, Cleroux et al., 2012). Although the temperature series from their youngest bivalve is about the same duration as our result (Figure 8), the two data sets are not directly comparable because the Wanamaker et al. (2007) series is annually resolved, colder than the open Gulf surface waters, and not likely to reflect only October SST. Nevertheless, each record captures ~2°C variability and the warming of recent decades.

5. CONCLUSIONS

We conclude that the Gulf of Maine, and specifically Jordan Basin, is a remarkable sediment archive of paleoclimate data that has been barely explored. Open marine (ice-free) sedimentation in Jordan Basin began about 16.9 kyr ago, with an abrupt change from red glacial marine sediments to gray-brown hemipelagic sediments. The deglacial to early Holocene rate of sedimentation is not well constrained in our study, but the late Holocene rate is about 330 cm per kyr on the basin floor. This means Jordan Basin sediments have the potential to record climate changes such as the North Atlantic Oscillation on decadal and longer time scales.

Our short (<1 year) sediment trap time series are the only data available on the seasonality of planktonic foraminiferal production from the Gulf of Maine. Combining the faunal data, an equally short series of oxygen isotope data from the same trap samples, and nearby hydrographic data indicates that *N. incompta* reached a flux maximum in late October 2010, and that its integrated $\delta^{18}\text{O}$ can be interpreted in terms of October sea surface temperature.

The sediment trap, hydrographic data, and climatological data have been used to develop a preliminary climate history of Jordan Basin based on radiocarbon dating of planktonic foraminifera and the $\delta^{18}\text{O}$ of *N. incompta*. We find that variability in sea surface temperature of about 2°C is typical for the Holocene, and that the two most recent cold periods (~1965, 1925) may be represented by maximum $\delta^{18}\text{O}$ in the upper 50 cm of the sediment column.

Calibration to modern sea surface conditions of proxy data, such as $\delta^{18}\text{O}$ measured in planktonic foraminifera, can best be done where core samples are located close to moored sediment trap and hydrographic stations. Jordan Basin in the Gulf of Maine is one of only a few such locations in the North Atlantic Ocean. As we rely more on paleo data to extend the short history of instrumental observations of climate, it will be increasingly important to have time-series sediment trap programs at locations that are critical to understanding climate and the ocean.

6. ACKNOWLEDGEMENTS

Cruise 198 of R/V Knorr was supported by the Grayce B. Kerr Fund. We are grateful to that fund and to the officers and crew of R/V Knorr. We appreciate the comments and advice of D. Belknap on Jordan Basin stratigraphy, and E. A. Boyle for sharing his unpublished Pb isotope results. M. Carman and E. Steponaitis processed sediment core samples for stable isotopes and radiocarbon, and K. Hayashi and E. Wood helped process sediment trap samples. A. Alpert extracted the historical SST data, and T. Bolmer prepared Figure 1. The manuscript was improved by the suggestions of three anonymous reviewers.

7. References

- Anand, P., Elderfield, H., and Conte, M., 2003, Calibration of Mg/Ca thermometry in planktonic foraminifera from a sediment trap time series: *Paleoceanography*, v. 18, no. 2, p. doi:10.1029/2002PA000846.
- Berger, W. H., and Killingley, J. S., 1977, Glacial-Holocene transition in deep-sea carbonates: selective dissolution and the stable isotope signal: *Science*, v. 197, p. 563-566.
- Bigelow, H. B., 1927, Physical Oceanography of the Gulf of Maine: *Bull. Bureau Fisheries*, v. 40, p. 511-1027.
- Black, D. E., Peterson, L. C., Overpeck, J. T., Kaplan, A., Evans, M. N., and Kashgarian, M., 1999, Eight Centuries of North Atlantic Ocean Atmosphere Variability: *Science*, v. 286, no. 26 November 1999, p. 1709-1713.
- Brown, W., and Irish, J., 1993, The annual variation of watermass structure in the Gulf of Maine: 1986-1987: *Journal of Marine Research*, v. 51, p. 53-107.
- Chapman, M., 2010, Seasonal production patterns of planktonic foraminifera in the NW Atlantic Ocean: Implications for paleotemperature and hydrographic reconstructions: *Paleoceanography*, v. 25, p. PA1101, doi:10.1029/2008PA0001708.
- Cleroux, C., Debret, M., Cortijo, E., Duplessy, J.-C., Dewilde, F., Reijmer, J., and Massei, N., 2012, High resolution sea surface reconstructions off Cape Hatteras over the last 10 ka: *Paleoceanography*, v. 27, no. PA1205, p. doi 10.1029/2011PA002184.
- Colton, J. B., 1968, Recent trends in subsurface temperatures in the Gulf of Maine and contiguous waters: *J. Fish Res. Bd. Canada*, v. 25, p. 2427-2437.
- Condrón, A., and Winsor, P., 2011, A subtropical fate awaited freshwater discharged from glacial Lake Agassiz: *Geophys. Res. Lett.*, v. 38, p. doi:10.1029/2010GL046011.
- Conte, M., Ralph, N., and Ross, E., 2001, Seasonal and interannual variability in deep ocean particle fluxes at the Oceanic Flux Program (OFP)/Bermuda Atlantic Time Series (BATS) site in the western Sargasso Sea near Bermuda: *Deep-Sea Research II*, v. 48, p. 1471-1505.
- Craig, H., and Gordon, L. I., 1965, Isotopic Oceanography: Deuterium and Oxygen 18 Variations in the Ocean and Marine Atmosphere, *in* Schink, D. R., and Corless, J. T., eds., *Symposium on Marine Geochemistry* Naragansett Marine Laboratory Occasional Publication #3, Univ. of Rhode Island, p. 277.
- Deuser, W. G., and Ross, E. H., 1989, Seasonally abundant planktonic foraminifera of the Sargasso Sea: Succession, deep-water fluxes, isotopic compositions, and paleoceanographic implications: *Journal of Foraminiferal Research*, v. 19, p. 268-293.
- Dickson, R. R., Meincke, J., Malmberg, S. A., and Lee, A. J., 1988, The "Great Salinity Anomaly" in the Northern North Atlantic 1968-1982: *Prog. Oceanog.*, v. 20, p. 103-151.
- Drinkwater, K., Belgrano, A., Borja, A., Conversi, A., Edwards, M., Greene, C., Ottersen, G., Pershing, A., and Walker, H., 2003, The response of marine ecosystems to

- climate variability associated with the North Atlantic Oscillation, *Geophysical Monograph 134*: Washington, DC, American Geophysical Union, p. 211-234.
- Durbin, E., Campbell, R., Casas, M., Ohman, M., Niehoff, B., Runge, J., and Wagner, M., 2003, Interannual variation in phytoplankton blooms and zooplankton productivity and abundance in the Gulf of Maine during winter: *Marine Ecology Progress Series*, v. 254, p. 81-100.
- Harbers, A., Schonfeld, J., Ruggeberg, A., and Pfannkuche, O., 2010, Short term dynamics of planktonic foraminiferal sedimentation in the Porcupine Seabight: *Micropaleontology*, v. 56, p. 259-274.
- Honjo, S., and Doherty, K., 1988, Large aperture time-series sediment traps; design objectives, construction and application: *Deep-Sea Research*, v. 35, p. 133-149.
- Houghton, R. W., and Fairbanks, R. G., 2001, Water sources for Georges Bank: *Deep-Sea Research II*, v. 48, p. 95-114.
- Hurrell, J. W., 1995, Decadal Trends in the North Atlantic Oscillation: Regional temperatures and precipitation: *Science*, v. 269, p. 676-679.
- Jonkers, L., Brummer, G.-J. A., Peeters, F., Aken, H. v., and Jong, M. F. D., 2010, Seasonal stratification, shell flux, and oxygen isotope dynamics of left-coiling *N. pachyderma* and *T. quinqueloba* in the western subpolar North Atlantic: *Paleoceanography*, v. 25, PA2204, doi:10.1029/2009PA001849.
- Kane, J., 2011, Multiyear variability of phytoplankton abundance in the Gulf of Maine: *Jour. Mar. Sci.*, v. 68, p. 1833-1841.
- Keigwin, L., Bice, M., and Copley, N., 2005, Seasonality and stable isotopes in planktonic foraminifera off Cape Cod, Massachusetts: *Paleoceanography*, v. 20, no. PA4011, doi:10.1029/2005PA001150.
- Keigwin, L. D., 1996, The Little Ice Age and Medieval Warm Period in the Sargasso Sea: *Science*, v. 274, p. 1504-1508.
- Lima, A. L., Bergquist, B. A., Boyle, E. A., Reuer, M. K., Dudas, F. O., Reddy, C. M., and Eglinton, T. I., 2005, High resolution historical records from Pettaquamscutt river basin sediments: 2. Pb isotopes reveal a potential new stratigraphic marker: *Geochim. et Cosmochim. Acta*, v. 69, p. 1813-1824.
- Mills, K., Pershing, A., Brown, C., Chen, Y., Chang, F.-S., Holland, D., Lehuta, S., Nye, J., Sun, J., Thomas, A., and Wahle, R., 2013, Fisheries management in a changing climate. Lessons from the 2012 ocean heat wave in the northwest Atlantic: *Oceanography*, v. 26, p. 191-195.
- Mountain, D. G., 2012, Labrador slope water entering the Gulf of Maine -response to the North Atlantic Oscillation: *Continental Shelf Res.*, v. 47, p. 150-155.
- Petrie, B., 2007, Does the North Atlantic Oscillation affect hydrographic properties on the Canadian Atlantic continental shelf?: *Atmosphere-Ocean*, v. 45, p. 141-151.
- Petrie, B., and Drinkwater, K., 1993, Temperature and salinity variability on the Scotian Shelf and in the Gulf of Maine 1945-1990: *Journal of Geophysical Research*, v. 98, p. 20,079-20,089.

- Pettigrew, N., Churchill, J., Janzen, C., Mangum, L., Signell, R., Thomas, A., Townsend, D., Wallinga, J., and Xue, H., 2005, The kinematic and hydrographic structure of the Gulf of Maine Coastal Current: *Deep-Sea Res. II*, v. 52, p. 2369-2391.
- Pilskaln, C., Anderson, D., McGillicuddy, D., Keafer, B., Hayashi, K., and Norton, K., 2013a, Spatial and temporal variability of Alexandrium cyst fluxes in the Gulf of Maine: Relationship to seasonal particle export and resuspension: *Deep-Sea Research II*, v. <http://dx.doi.org/10.1016/j.dsr2.2012.11.001i>.
- Pilskaln, C., Hayashi, K., Keafer, B., Anderson, D., and McGillicuddy, D., 2013b, Benthic nepheloid layers in the Gulf of Maine and Alexandrium cyst inventories: *Deep-Sea Research II*, v. <http://dx.doi.org/10.1016/j.dsr2.2013.05.021i>.
- Pilskaln, C., Manganini, S., Trull, T., Armand, L., Howard, W., Asper, V., and Massom, R., 2004, Geochemical particle fluxes in the Southern Indian Ocean seasonal ice zone: Prydz Bay region, East Antarctica: *Deep-Sea Research I*, v. 51, p. 307-332.
- Pilskaln, C., Paduan, J., Chavez, F., Anderson, R., and Berelson, W., 1996, Carbon export and regeneration in the coastal upwelling system of Monterey Bay, central California: *Jour. Mar. Res.*, v. 54, p. 1149-1178.
- Pilskaln, C. H., and Paduan, J. B., 1992, Laboratory techniques for the handling and geochemical analysis of water column particulate and surface sediment samples, Volume Technical Report 92-9: Monterey, CA, Monterey Bay Aquarium Research Institute (MBARI), p. 22p.
- Schnitker, D., Belknap, D. F., Bacchus, T. S., Friez, J. K., Lusardi, B. A., and Popek, D. M., 2001, Deglaciation of the Gulf of Maine, in Weddell, T., and Retelle, M., eds., *Deglacial history and relative sea-level changes, northern New England and adjacent Canada*, Volume Special Paper 351: Boulder, CO, Geol. Soc. Amer.
- Shackleton, N. J., 1974, Attainment of isotopic equilibrium between ocean water and the benthonic foraminifera genus *Uvigerina*: isotopic changes in the ocean during the last glacial: *CNRS Colloque*, v. 219, p. 203-209.
- Stuiver, M., and Reimer, P. J., 1993, Extended ¹⁴C database and revised CALIB radiocarbon calibration program: *Radiocarbon*, v. 35, p. 215-230.
- Thomas, E., Wolff, E., Mulvaney, R., Steffensen, J., Johnsen, S., Arrowsmith, C., White, J., Vaughn, B., and Popp, T., 2007, The 8.2 ka event from Greenland ice cores: *Quat. Sci. Rev.*, v. 26, p. 70-81.
- Todd, B., and Shaw, J., 2012, Laurentide Ice Sheet dynamics in the Bay of Fundy, Canada, revealed through multibeam sonar mapping of glacial landsystems: *Quat. Sci. Rev.*, v. 58, p. 83-103.
- Townsend, D. W., Rebeck, N. D., Thomas, M. A., Karp-Boss, L., and Gettings, R. M., 2010, A changing nutrient regime in the Gulf of Maine: *Continental Shelf Res.*, v. 30, p. 820-832.
- Tucholke, B., and Hollister, C., 1973, Late Wisconsin glaciation of the southwestern Gulf of Maine: New evidence from the marine environment: *Geological Society of America Bulletin*, v. 84, p. 3279-3296.
- Wanamaker, A. D., Kreutz, K. J., Schone, B. R., Pettigrew, N., Borns, H. W., Introne, D. S., Belknap, D., Maasch, K. A., and Feindel, S., 2007, coupled North Atlantic

slope water forcing on Gulf of Maine temperatures over the past millennium:
Clim. Dyn., v. DOI 10.1007/s00382-007-0344-.