Deglacial Floods in the Beaufort Sea Preceded Younger Dryas Cooling

L.D. Keigwin[1*], S. Klotsko[2], N. Zhao[1], B. Reilly[3], L. Giosan[1], and N.W. Driscoll[2]

[1]Woods Hole Oceanographic Institution, Woods Hole, MA 02543.
[2]Scripps Institution of Oceanography, UCSD, La Jolla, CA 92093.
[3]College of Earth, Ocean, and Atmospheric Sciences, OSU, Corvallis, OR 97331.
*Correspondence to: lkeigwin@whoi.edu

The Younger Dryas cooling at ~13 ka, after 2 kyr of postglacial warming, is a century-old climate problem. The Younger Dryas is thought to have resulted from a slow-down of the Atlantic meridional overturning circulation in response to a sudden flood of Laurentide Ice Sheet meltwater that reached the Nordic Seas. Although there is no oxygen isotope evidence in planktonic foraminifera from the open western North Atlantic for a local source of meltwater from the Gulf of St. Lawrence where it was predicted, we report here that the eastern Beaufort Sea contains the long-sought signal of ^{18}O-depleted water. Beginning at ~12.94 ± 0.15 ka, oxygen isotopes in planktonic foraminifera from two sediment cores as well as sediment and seismic data indicate a flood of melt water, ice and sediment to the Arctic via Mackenzie River that lasted about 700 years. The minimum in oxygen isotope ratios lasted ~130 years. The floodwater would have travelled north along the Canadian Archipelago, and through Fram Strait to the Nordic Seas where freshening and freezing near sites of deepwater formation would have suppressed convection, and caused the Younger Dryas cooling by reducing the meridional overturning.

Introduction

It is known that conditions in the Arctic Ocean have a profound effect on the North Atlantic Ocean, for example the Great Salinity Anomaly (GSA) of the 1960s and 1970s[1] and that export of excess fresh water and ice through Fram Strait was the origin of the GSA[2,3]. During transit of the GSA around convective regions of the Nordic Seas, decreased sea surface salinities and increased sea ice cover reduced convective overturn and contributed to very harsh winters. There is reason to expect that similar and even larger climate events occurred in the past, especially during deglaciation when huge volumes of meltwater and ice suddenly entered the Gulf of Mexico, the Arctic, and Nordic Seas. For example, it was discovered several decades ago that an abrupt decrease in the oxygen isotope ratio ($\delta^{18}O$) in surface-dwelling planktonic foraminifera midway through deglaciation in the Gulf of Mexico was a signal of a fresh water flood[4]. The source of this runoff must have been the decaying Laurentide Ice Sheet (LIS) via the Mississippi River, but it ended abruptly at about 13 ka[5]. Kennett and Shackleton[4] proposed that, as the southern margin of the LIS retreated northward, meltwater was routed eastward to the Gulf of St. Lawrence, and the western North Atlantic. The $\delta^{18}O$ decrease in the Gulf of Mexico was more than 2 ‰, so a signal of 1 ‰ or more should stand out in fresher, higher latitude waters. However, a low $\delta^{18}O$ signal at about 13 ka has never been detected in high-quality sediment cores from the open western North Atlantic[6,7,8], yet it is believed that the diversion of the flood from the Gulf of Mexico interrupted deep ocean convection and caused the well-known Younger Dryas (YD) cold episode (12.9-11.7 ka) in the North Atlantic region[9]. Other proposed explanations for the origin of the YD include melting of the Fennoscandian ice sheet [10], changes in atmospheric circulation[11], and a combination of ice sheet melting, atmospheric flow patterns, and radiative forcing[12]. However, those explanations beg the question where did the diverted meltwater go.

The YD was discovered near the beginning of the 20th Century as one of several appearances of the Arctic wildflower *Dryas octopetala* in postglacial deposits in Scandinavia[13,14] and eventually was defined as a useful chronostratigraphic zone in the North Atlantic region[15]. It was later proposed that meltwater routing and drainage pattern changes could have caused the YD by lowering surface ocean salinity[16,17].

Recently, a glacial systems model showed that fresh water stored in glacial Lake Agassiz most likely traveled north to the Beaufort Sea via Mackenzie River at 13 ka[18], and extensive field work on the Mackenzie Delta identified clear evidence of massive flood deposits that occurred about the same time[19]. Although the exact timing and magnitude of the Murton et al.[19] conclusions have been questioned[20], application of a high resolution ocean circulation model[21] showed that, only when released to the Arctic Ocean (via Mackenzie River), could the Lake Agassiz flood have caused the Younger Dryas reduction of Atlantic meridional overturning circulation (AMOC)[22] and consequent northern hemisphere cooling.

Results and Discussion

Here we present data that show two events of substantial sea surface freshening during deglaciation in newly acquired large diameter (jumbo) piston cores (JPCs) from 690 m on the continental slope ~100 km east of Mackenzie River (JPCs 15 and 27, Fig. 1). These and other new cores underlie the Atlantic water that enters the Arctic at Fram Strait and the Barents Sea in the depth range ~100 to 800 m and circulates counter-clockwise along the continental slopes[23]. JPC-15 penetrated ~13 m of sediment and was probably stopped by a coarse ice rafted debris (IRD) layer that has high magnetic susceptibility, high Ca content (Fig. 2), and makes a prominent reflector in the acoustic stratigraphy across the region (Fig. 3). Later, reoccupying the same site, a longer (heavier) deployment (JPC-27) penetrated the deeper coarse layer. As these cores have nearly identical lithology, we spliced them together to make a composite JPC-15/27 (Fig. Extended Data (ED)1).

Compared to the lower layer, the upper coarse layer at this site is thicker, has multiple events, and fewer IRD grains (Fig. 2), but each layer also has finer sand and silt (Fig. 3). These data indicate that each coarse layer provides a record of enhanced sediment transport to the upper slope of the Beaufort Sea. The two main events must be the same that Scott et al.[24] noted in Canadian core PC-750 (Fig. 1). X-ray fluorescence (XRF) counts of calcium (interpreted as detrital $CaCO_3$ content) show that our two events have similar carbonate content, but also that lowest carbonate delivery to the region occurred before the oldest event and was only a little higher between the events (Fig. 2). Sediment deeper than ~5 m is faintly laminated at the cm scale, except for the massive appearance of the first event (13.0-13.5 m). Laminae are better developed between 6 and 12 m, where about 300 were counted in XRF data (Fig. ED2).

As with the sediment and geophysical data, $\delta^{18}O$ on the polar planktonic foraminifer *Neogloboquadrina pachyderma* (Ehrenberg) left coiling (Nps) in JPC-15/27 is marked by two prominent events at the same depths in the core (Fig. 2). At those levels $\delta^{18}O$ of Nps ($\delta^{18}O_{Nps}$) decreased at least 1.0 ‰ below the ~2.0 ‰ baseline that extends >4 m down the core. Above a pronounced maximum in $\delta^{18}O_{Nps}$ at 2-3 m, values decrease by ~2.0 ‰ to the core top. We find the upper $\delta^{18}O_{Nps}$ minimum to the west of JPC15/27 at JPC-09 (see ED), but not at other cores farther west (Figure 1). Benthic foraminiferal (*Cassidulina neoteretis* Seidenkrantz) $\delta^{18}O$ at JPC15/27 yields a stratigraphy more typical of the world ocean, with generally increasing values down the core, although they are consistently low at about 5 m in the same samples where $\delta^{18}O_{Nps}$ is low (Fig. 2E).

Chronology

Chronology in Arctic sediments is uncertain because, although radiocarbon dating of foraminifera is the simplest method, at present there is no way to know exactly the near surface reservoir correction (ΔR) in the past. We made 14 accelerator mass spectrometer (AMS) ^{14}C measurements on Nps from core 15 (Table ED1). Those dates indicate maximum rates of sedimentation of at least 10 m per 1000 ^{14}C yrs between 6 and 12 m in the mid-deglacial interval of the core (Fig. 2F, ED3). Before and after that mid-core extreme, rates are half that or less. We assume Nps calcified near the interface between fresher, near-surface water and underlying saltier water, as it does today[25]. In that environment it was salty enough to survive but shallow enough to capture low $\delta^{18}O$ events. Six of the Nps dates were paired with dates on *C. neoteretis.* On average, benthics are only 120 ±220 yrs older than planktonics, including a result from 1300 m in the Chukchi Sea (HLY0205 JPC-16, Fig. 1)[26]. This small difference indicates upper slope waters were relatively well ventilated.

For a calendar (calibrated) ^{14}C chronology, we need to choose a ΔR. In the modern Arctic the Pacific inflow through Bering Strait is a source of old carbon that would have been absent prior to about 11 ka when the strait was dry land[26,27] (see Fig. ED4 and further discussion). Bondevik et al.[28] showed that surface waters along the Norwegian coast, which would have fed the Arctic, had a ΔR of about 0 yrs, like today, during the Bølling-Allerød (B/A), about 100 yrs early in the YD, and about 200 yrs during the mid-YD. Cao et al.[29] reached a similar conclusion based on U series dates and ^{14}C measurements on a solitary coral from the southern Labrador Sea, which would monitor intermediate depth waters leaving the Nordic Seas. Therefore, we developed an age model using a Bayesian method and $\Delta R = 0 \pm 100$ yrs for the Holocene and the Allerød, and 200 ± 100 yrs for the YD (Fig. ED5). If the relatively high ΔR during the YD was triggered by an event late in the Allerød, then the onset of that event should be calibrated with a ΔR of ~0 years

These estimates of only a modest ΔR (0-200 years) encompass the pre-bomb estimate[30] based on ^{14}C, tritium, and $\delta^{18}O$ on samples collected decades ago when bomb produced nuclides were beginning to invade the deep Arctic. Ostlund et al.[30] inferred the pre-bomb ^{14}C activity of waters between 500 and 1500 m was about -55 ± 5 ‰, giving a ΔR of ~40 yrs (Fig. 4). Therefore, although deep-water ^{14}C circulation in the Arctic may have been very different in the past[31,32], it appears that the ^{14}C ventilation of upper waters (<1500 m) in the Canada Basin was similar to today.

Our age model gives interpolated calendar ages in JPC-15/27 of 12.94 ± 0.15 ka for the onset of the upper $\delta^{18}O$ minimum (constrained by dates of 13.06 ka at 520 cm and 12.66 ka at 500 cm; Table ED1), and ~14.6 ka for the peak of the older one (Fig. 5). The age of the older event and its associated IRD is consistent with the ages (15.2-14.1 ka) reported for the initial withdrawal of ice streams from Amundsen Gulf and M'Clure Strait[33]. The age for the onset of the later freshening in the Beaufort Sea (12.94 ± 0.15 ka) is virtually the same as the beginning of the Younger Dryas at 12.85 ±0.14 ka in Greenland ice cores[34], and identical to the end of freshening in the Gulf of Mexico (12.94 ±0.17 ka) (Fig. 5). Minimum $\delta^{18}O$ in JPC-15/27 first occurred at 12.59 ± 0.14 ka, within uncertainty of the equivalent event at JPC-09 (12.7 ± 0.10 ka). The coincidence of the end of flooding in the Gulf of Mexico and the onset of flooding in the eastern Beaufort Sea is strongly suggestive that the routing of meltwaters[35] switched from the Gulf of Mexico to the Beaufort Sea at about 13 ka.

Ocean and climate change in Beaufort Sea

Our composite sequence from the continental slope east of Mackenzie River began around 15-16 ka with modest ice rafting from local sources such as ice streams in M'Clure Strait and Amundsen Gulf. Icebergs would have traveled clockwise around Canada Basin via the Beaufort Gyre, and the counter clockwise shelfbreak current[36] would have been weakened with sea level below the depth of Bering Strait. Mackenzie River may not have supplied substantial detrital carbonate because the extensive Devonian carbonate terrain south of Great Slave Lake and north of Ft. McMurray[37] was probably ice-covered, but it may have been a source of runoff and sediment at least since ~18 ka based on the background low $\delta^{18}O$ (Fig. 5). This was a time when secular change in the ocean due to increased ice volume was about +1 ‰, indicating the sea surface was less saline than today by at least 1 psu, assuming the modern $\delta^{18}O$-salinity relationship[38]. This setting prevailed until ~14.6 ka, when ice rafting dramatically increased from Amundsen Gulf and M'Clure Strait, and $\delta^{18}O_{Nps}$ decreased by >1 ‰. The five samples defining this minimum were probably deposited within decades.

At the end of the 14.6 ka event, Amundsen Gulf probably remained a source of sediment to the continental slope in the eastern Beaufort Sea until the ice stream was fully retreated. Mackenzie River may have always been a large source of sediment, but as more of its watershed north of Fort McMurray was deglaciated, the more important it must have become. The laminated sediments, high sedimentation rate, and general lack of coarse particle ice rafting suggest large sediment input from the Mackenzie River between 13.5 and ~14.4 ka (6-12 m in the core). The high sedimentation rates along the slope may be explained by discharge over bottom fast ice on the shelf, which could efficiently transport sediment farther seaward (e.g., ref 39). Based on the diagnostic acoustic signature of the rapidly emplaced Bølling-Allerød section (Fig. 3), the western extent of the deposit pinches out between JPC-09 and JPC-06 (Fig. 1, Fig. ED8). Counts of ~300 layers within the ~900 year interval where sedimentation rates are highest show the layers are probably not annual (Fig. ED2). The interval between the two $\delta^{18}O_{Nps}$ minima represents most of the Bølling-Allerød climate warming, when the AMOC was almost as strong as today[22], but evidently the lowered salinity in the Beaufort Gyre had little direct influence on North Atlantic overturning. This may indicate that the gyre was in a mostly anticyclonic state, which today stores ice and fresh water[40].

Close to 13 ka the rapid increase in magnetic susceptibility and decreased $\delta^{18}O_{Nps}$ in JPC-15/27 herald the beginning of the YD. Although the two $\delta^{18}O_{Nps}$ minima in this core are similar in size, the YD event was more likely to have been a flood of fresh water with high suspended load[19] because $\delta^{18}O$ of *C. neoteretis*, living at the seafloor, decreased in exactly the same samples as Nps during the ~13 ka but not the earlier event. This, we propose, may record a hyperpycnal flow that brought low salinity to the seafloor and that would be more likely from a river flood. The major sediment depocenter in this model must be farther seaward because sedimentation rates drop during this time interval at 690 m water depth (Fig. 2). The YD flood can be traced to the west as far as core JPC-09 using $\delta^{18}O_{Nps}$, but the signal is not clear west of that at JPC-06, and neither the $\delta^{18}O$

minima nor the maximum in magnetic susceptibility are evident as far west as JPC-02 near Barrow, Alaska (Fig. ED7).

About 200 yrs after the onset of the YD flood all four sediment and isotope proxies were briefly aligned in the first (labeled “a”) of several sub-events (Fig. 2 A-D). The low $\delta^{18}O_{Nps}$ episode is mostly centered between the subpeaks “a” and “b” of the magnetic susceptibility (12.8 to 12.3 ka), but the last of the spikes in IRD and carbonate deposition ended with increased $\delta^{18}O_{Nps}$ at the end of flooding. Maximum $\delta^{18}O_{Nps}$ at ~12.2 ka probably marks an interval of relatively high salinity in the near surface Beaufort Sea[41], followed by more typical decreasing $\delta^{18}O$ trends in benthic and planktonic foraminifera as ice volume decreased and climate warmed during the Holocene. The lingering high magnetic susceptibility late in the YD may indicate evolving sources of sediment from Mackenzie River, and it might also relate to evidence of a second flood ~11 ka (ref. 19).

Knowing the duration of the YD flood is important for calculating the fresh water transport and evaluating its effect on the AMOC. If we take the main flood interval of the YD as that part where $\delta^{18}O_{Nps}$ was less than the 2 ‰ baseline, then it lasted ~660 years. If the lowest $\delta^{18}O_{Nps}$ indicates peak discharge, then most of the fresh water transport could have occurred in about 130 years (Table ED3). However, it must be kept in mind that if the Mackenzie River choke point at Fort McMurray was breached suddenly at the beginning of the YD, and this is contentious[42], then the outburst of Glacial Lake Agassiz water would have probably produced initial salinities over our core site that were too low for Nps to grow. Furthermore, estimates of very high fresh water transport during the flood are based on the assumption that it occurred on the timescale of a year[43], yet if the main flood was that brief then it is unlikely enough planktonic foraminifera could have recorded the low $\delta^{18}O$ to leave a signal in the geological record.

Most likely the initial Mackenzie discharge at 12.9 ka was a combination of both a routing change from the Gulf of Mexico and an outburst flood from glacial Lake Agassiz. This potent combination of two sources of fresh water was probably effective in reducing the AMOC[35], especially considering it was an Arctic source[21]. However, even if the combined routing plus glacial lake release to Mackenzie River itself was too modest to trigger a collapse of the AMOC, many large rivers empty into the Arctic[2], and Lena River, one of the largest, also flooded about 13 ka (ref 44). Finally, it should be noted that in addition to fresh water floods in the Arctic around the beginning of the YD, it is reported that enhanced sea ice export through Fram Strait at that time also had a Beaufort Sea source[45].

By the onset of the YD, the AMOC may have already been close to a tipping point after ~1500 years of low salinity leakage from the Beaufort Sea, and transport to the nearshore convective regions of the Nordic seas[46,47]. Increased freshening has also been noted at other coastal locations in the North Atlantic, including the proposed eastern outlet (St. Lawrence River system) using various proxies[48,49,50], the Baltic Sea[10, 51], and off eastern Greenland where $\delta^{18}O_{Nps}$ minima of YD age are thought to reflect local melting[52] but could also be evidence of the Mackenzie flood. The coincidence of decreased $\delta^{18}O$ in the Beaufort Sea and increased $\delta^{18}O$ in the Gulf of Mexico near the beginning of the YD is a good test of the meltwater diversion hypothesis of Kennett and Shackleton[4] (Fig. 5). Considering all the other observations, including the climatic background suggested by alternative hypotheses[10-12] that may have helped sustained the event, and the lack of a large YD minimum in $\delta^{18}O$ anywhere in the open North Atlantic Ocean, the ~12.9 ka flood of Mackenzie River was most likely the trigger for the reduction of the AMOC and Younger Dryas cooling.

References

1. Dickson, R. R., Meincke, J., Malmberg, S. A. & Lee, A. J. The “Great Salinity Anomaly" in the Northern North Atlantic 1968-1982. *Prog. Oceanog.* **20**, 103-151 (1988).

2. Aagard, K. & Carmack, E. The role of sea ice and other fresh water in the Arctic circulation. *Jour. Geophys. Res.* **94**, 14485-14498 (1989).

3. Häkkinen, S. An Arctic source for the great salinity anomaly: A simulation of the Arctic ice-ocean system for 1955-1975. *Jour. of Geophys. Res.* **98**, 16,397-316,410 (1993).

4. Kennett, J. P. & Shackleton, N. J. Laurentide ice sheet meltwater recorded in Gulf of Mexico deep-sea cores. *Science* **188**, 147-150 (1975).

5. Williams, C., Flower, B. & Hastings, D. W. Seasonal Laurentide ice sheet melting during the "Mystery Interval" (17.5-14.5 ka). *Geology* **40**, 955-958 (2012).

6. Keigwin, L. D. & Jones G. A. The marine record of deglaciation from the continental margin off Nova Scotia. *Paleoceanography* **10**, 973-985 (1995).

7. de Vernal, A., Hillaire-Marcel, C. & Bilodeau, G. Reduced meltwater outflow from the Laurentide ice margin during the Younger Dryas. *Nature* **381**, 774-777 (1996).

8. Keigwin, L. D., Sachs, J., Rosenthal, Y. & Boyle, E. A. The 8200 year B.P. event in the slope water system, western subpolar North Atlantic. *Paleoceanography* **20**, doi:10.1029/2004PA001074 (2005).

9. Broecker, W. S., et al. The routing of meltwater from the Laurentide ice-sheet during the Younger Dryas cold episode. Nature **341**, 318-321 (1989).

10. Muschitiello, F. et al. Fennoscandian freshwater control on Greenland hydroclimate shifts at the onset of the Younger Dryas. *Nature Comm*. (2015).

11. Brauer, A. Haug, G.H., Dulski, P., Sigman, D. M., & Negendank, J. F. W. An abrupt wind shift in western Europe at the onset of the Younger Dryas cold period. *Nature Geoscience* **1**:520-523 (2008).

12. Renssen, H. et al. Multiple causes of the Younger Dryas cold period. *Nature Geoscience* **8**, 946-950 (2015).

13. Andersson, G. Swedish vegetation history. Stockholm, P.A. Norstedt & Soners Forlag (1897).

14. Hartz, N. & Milthers, V. The late glacial clay in the Allerod brickyard. *Meddelelser Dansk Geologisk Foreningen* **8**, 31-60 (1901).

15. Mangerud, J., Andersen, S. T., Berglund, B. E., & Donner, J. J. Quaternary stratigraphy of Norden, a proposal for terminology and classification. *Boreas* **3**, 110-127 (1974).

16. Johnson, R. G. & McClure B. T. A model for northern hemisphere continental ice sheet variation. *Quaternary Res.* **6**, 325-353 (1976).

17. Rooth, C. Hydrology and ocean circulation. *Prog. Oceanog.* **11**, 131-149 (1982).

18. Tarasov, L. & Peltier, W. R. Arctic freshwater forcing of the Younger Dryas cold reversal. *Nature* **435**, 662-665 (2005).

19. Murton, J. B., Bateman, M. D., Dallimore, S. R., Teller, J. T. & Yang, Z. Identification of Younger Dryas outburst flood path from lake Agassiz to the Arctic Ocean. *Nature* **464**, 740-743 (2010).

20. Carlson, A.E. et al. Geochemical proxies of North American freshwater routing during the Younger Dryas cold event. *Proc. Nat. Acad. Sci.* **104**, 6556-6561 (2007).

21. Condron, A. & Winsor, P. Meltwater routing and the Younger Dryas. *Proc. Nat. Acad. Sci.* doi:10.1073/pnas.1207381109 (2012).

22. McManus, J. F., Francois, R., Gherardi, J.-M., Keigwin, L. D. & Brown-Leger, S. Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes. *Nature* **428**, 834-837 (2004).

23. Rudels, B., Jones, E. P., Anderson, L. G. & Kaattner, G. On the intermediate depth waters of the Arctic Ocean. *Geophysical Monograph* **85**, 33-46 (1994).

24. Scott, D., Schell, T., St-Onge, G., Rochon, A. & Blasco, S. Foraminiferal assemblage changes over the last 15,000 years on the Mackenzie-Beaufort Sea slope and Amundsen Gulf, Canada: Implications for past sea ice conditions. *Paleoceanography* **24** (doi:10.1029/2007PA001575) (2009).

25. Bauch, D., Carstens, J, & Wefer, G. Oxygen isotope composition of living *Neogloboquadrina pachyderma* (sin.) in the Arctic Ocean. *Earth Planet. Sci. Lett.* **146**, 47-58 (1997).

26. Keigwin, L. D, Donnelly, J. P., Cook, M.S., Driscoll, N.W. & Brigham-Grette, J. Flooding of Bering Strait and Holocene climate in the Chukchi Sea. *Geology* **34**, 861-864 (2006).

27. Jakobsson, M. et al. Post-glacial flooding of the Beringia Land Bridge dated to 11,000 cal yrs BP based on new geophysical and sediment records. *Clim. Past Discuss*., doi:10.5194/cp-2017-11 (2017).

28. Bondevik, S., Mangerud, J., Birks, H. H., Gulliksen, S. & Reimer, P. Changes in North Atlantic radiocarbon reservoir ages during the Allerod and Younger Dryas. Science **312**, 1514-1517 (2006)

29. Cao, L., Fairbanks, R.G., Mortlock, R.A. & Risk M.A. Radiocarbon reservoir age of high latitude North Atlantic surface water during the last deglacial. *Quat. Sci. Rev*. **26**, 732-742 (2007).

30. Ostlund, H., Possnert, G. & Swift, J. Ventilation rate of the deep Arctic Ocean from carbon 14 data. *Jour. Geophys. Res*. **92**, 3769-3777 (1987).

31. Cronin, T. et al. Deep Arctic Ocean warming during the last glacial cycle. *Nature Geoscience* **5**, 631-634 (2012).

32. Thornalley, D. J. R. et al. A warm and poorly ventilated deep Arctic Mediterranean during the last glacial period. *Science* **349,** 706-710 (2015).

33. Stokes, C. R., Clark, C. D. & Storrar, R. Major changes in ice stream dynamics during deglaciation of the north-western margin of the Laurentide ice sheet. *Quat. Sci. Rev.* **28**, 721-738 (2009).

34. Rasmussen, S., et al. A new Greenland ice core chronology for the last glacial termination. *Jour. Geophys. Res.* **111**, doi:10.1029/2005JD006079 (2006).

35. Meissner, K. & Clark, P. Impact of floods versus routing events on the thermohaline circulation. *Geophys. Res. Lett.* **33,** doi:10.1029/2006GL026705 (2006).

36. von Appen, W.-J. & Pickart, R. Two configurations of the western Arctic Shelfbreak Current in summer. *Jour. Phys. Oceanography* **42**, 329-351 (2012).

37. Wheeler, J., et al. Geological Map of Canada. "A" Series Map 1860A; doi:10.4095/208175, Geological Survey of Canada (1996).

38. Cooper, L., et al. Linkages among runoff, dissolved organic carbon, and the stable oxygen isotope composition of seawater and other water mass indicators in the Arctic Ocean. *Jour. Geophys. Res.* **110,** (doi:10.1029/2005JG000031) (2005).

39. Macdonald, R. & Yu, Y. The Mackenzie estuary of the Arctic OCean. *Handbook of Environmental Chemistry.* Berlin, Springer-Verlag. **5:** 91-120 (2006).

40. Proshutinsky, A. & Johnson, M.A. Two circulation regimes of the wind-driven Arctic Ocean. *Jour. Geophys. Res.* **102**,12493-12514 (1997).

41. Schell, T., Scott, D. B., Rochon, A. & Blasco, S. Late Quaternary paleoceanography and paleo-sea ice conditions in the Mackenzie Trough and Canyon, Beaufort Sea. *Can. J. Earth Sci*. **45**, 1399-1415 (2008).

42. Fisher, T.G., Waterson, N., Lowell, T. V. & Hajdas, I. Deglaciation ages and meltwater routing in the Fort McMurray region, northeastern Alberta and northwestern Saskatchewan, Canada. *Quat. Sci. Rev.* **28**, 1608-1624 (2009).

43. Leverington, D. W., Mann, J. D. & Teller, J. T. Changes in the bathymetry and volume of glacial Lake Agassiz between 11,000 and 9300 ^{14}C yr B.P. *Quaternary. Res.* **54**, 174-181 (2000).

44. Spielhagen, R.F., Erlenkeuser, H. & Siegert, C. History of freshwater runoff across the Laptev Sea (Arctic) during the last deglaciation. *Global and Planetary Change* **48**, 187-207 (2005).

45. Hillaire-Marcel, C., Maccali, J., Not, C. & Poirier, A. Geochemical and isotopic tracers of Arctic sea ice sources and export with special attentiion to the Younger Dryas interval. *Quat. Sci. Rev.* **79**, 184-190 (2013).

46. Mauritzen, C. Production of dense overflow waters feeding the North Atlantic across the Greenland-Scotland Ridge. Part 1: Evidence for a revised circulation scheme. *Deep-Sea Res.* **43**, 769-806 (1996).

47. Pedlosky, J. & Spall, M. Boundary intensification of vertical velocity in a beta-plane basin. *Jour. Phys. Oceanography* **35**, 2487-2500 (2005).

48. Carlson, A.E. & Clark, P.U. Ice sheet sources of sea level rise and freshwater discharge during the last deglaciation. *Rev. Geophys*, **50** (2011RG000371) (2012).

49. Cronin, T.M., Rayburn, J.A., Guilbault, J.-P. & Thunell, R. Stable isotope evidence for glacial lake drainage through the St. Lawerence estaury, eastern Canada, ~13.1-12.9 Ka. *Quat. International* **260**, 55-65 (2012).

50. Levac, E., Lewis, M., Stretch, V., Duchesne, K. & Neulieb, T. Evidence for meltwater drainage via the St. Lawrence River valley in marine cores from

the Laurentian Channel at the time of the Younger Dryas. *Global and Planetary Change* **130**, 47-65 (2015).

51. Boden, P., Fairbanks, R.G., Wright, J. D. & Burckle, L. H. High-resolution stble isotope records from southwest Sweden: The drainage of the Baltic Ice Lake and Younger Dryas ice margin oscillations. *Paleoceanography* **12**, 39-49 (1997).

52. Jennings, A.E, Hald, M., Smith, M. & Andrews, J.T. Freshwater forcing from the Greenland Ice Sheet during the Younger Dryas: evidence from southeastern Greenland shelf cores. *Quat. Sci. Rev.* **25**, 282-298 (2006).

53. Jakobsson, M. et al. The International bathymetric chart of the Arctic Ocean (IBCAO) version 3.0. *Geophys. Res. Lett*. Doi: 10.1029/2012GL052219 (2012).

54. Leventer, A., Williams, D. F. & Kennett, J. P. Dynamics of the Laurentide ice sheet during the last deglaciation: evidence from the Gulf of Mexico. *Earth and Planet. Sci. Lett.* **59**,11-17 (1982).

Acknowledgements We thank the officers and crew of USCGC Healy for making this project a success. We are also indebted to M. Carman for help processing core samples; A. McNichol for helpful discussions of ^{14}C in the Arctic; A. Gagnon for the stable isotope measurements; the NOSAMS staff for providing ^{14}C data; M. McCarthy, C. Moser, C. Griner, and C. Mayo for leading the coring effort. S. Nielsen helped with translation of ref. 14. The manuscript benefited from comments of 4 anonymous reviewers and help from M. Blaauw and B. Keigwin with the Bayesian age model. This research was funded by NSF grants ARC 1204045 to L.D.K., and ARC 1203944 to N.W.D.

Author Contributions L.D.K. conceived the project, N.W.D. and S.K surveyed the seafloor, identified coring locations, and studied grain size; B.R. conducted the magnetic measurements; and N.Z. and L.G. conducted the XRF scanning. All authors helped write the manuscript.

Author Information Reprints and permissions information is available at www. nature.com/reprints. The authors have no competing financial interests. Correspondence and requests for materials should be addressed to lkeigwin@whoi.edu

Methods

Site survey methods, laminae counting methods, core sampling and stable isotope methods, sample preparation for radiocarbon dating, and Bayesian age modeling are presented in Supplementary Information sections 1, 2, 3, 4.2, and 4.4.

Data Availability:

Radiocarbon data appear in Extended Data Table 1, and stable isotope data appear in Extended Data Table 3. The Fe/Ca and magnetic data are available from the authors upon request.

Code availability:

The code for the bacon age model is freely available; see http://chrono.qub.ac.uk/blaauw/bacon.html. The settings used are found in Supplementary Section 4.4.

Figure Captions

Fig. 1. Overview of core locations and stratigraphy in the eastern Beaufort Sea. (Top) Track of USCGC Healy cruise 1302 with location of JPC sites (yellow) and other cores (red) discussed in this study. Inset (based on ref 27) shows the study area with respect to the Arctic Ocean. LR=Lomonosov Ridge, CB=Chukchi Borderland, BC=Barrow Canyon, MT=Mackenzie Trough, AG=Amundsen Gulf, MS=M'Clure Strait, BI=Banks Island, VI=Victoria Island. Based on low $\delta^{18}O_{Nps}$ and seismic evidence, the YD flood deposit (sites in yellow with a cross) ranges from JPC-09 in the west to JPC-22 in the east. Because of Coriolis force and lowered sea level, the flood would have travelled north and east. (Bottom) Down core magnetic susceptibility is shown delineating the Holocene (yellow) - Deglacial (blue) boundary. Selected AMS ^{14}C dates are calibrated ka.

Fig. 2. Proxy data from JPC-15/27 in the eastern Beaufort Sea. Magnetic susceptibility (A), lithic particle abundance (B), Ca content (proxy for $CaCO_3$) (C), and $\delta^{18}O_{Nps}$ (D) all exhibit extreme values early the Bolling/Allerod warming at 14.6 ka (red line at 1300 cm) and during the YD (11.7-12.9 ka) (~380-510 cm). Dashed vertical lines correlate smaller features. Dashed horizontal line in (D) is a ~2.0 ‰ reference for $\delta^{18}O$ features. Data corresponding to a large dropstone at 1346-1355 cm excluded from C. The *C. neoteretis* (benthic) $\delta^{18}O$ (E) is unremarkable except that the clear minimum ~450-500 cm occurs in the same samples as the low $\delta^{18}O_{Nps}$. Sedimentation rates (F) are very high where sediments are laminated, although ^{14}C dates exaggerate the B/A maximima (Figure ED2),.

Fig. 3. Grain size variability down composite jumbo piston core JPC15/27. Magnetic susceptibility data are superposed on the grain size and the seismic data, assuming the pressure wave velocity of the core logged at sea (1333 ms^{-1}). These properties all vary together. The seismic data show a diagnostic reflector pattern with an upper (~380-520 cm) and lower high (~1320 cm) amplitude reflectors that bound a region of lower acoustic reflectivity. The zone of lower reflectivity correlates with high sediment accumulation rates, low magnetic susceptibility, low ice rafted debris (IRD), and low Ca content (Fig. 2).

Fig. 4. Radiocarbon basis for the age model in this paper. Ostlund et al.[30] synthesized $\Delta^{14}C$, $\delta^{18}O$, and tritium data collected from several Arctic ice camps (LOREX, CESAR, AIWEX) between 1977 and 1985 and concluded that the pre-bomb value of intermediate depth waters (500 to 1500 m) was -55 ±5 ‰ (vertical black line $\pm 1\sigma$ (dashed)), and pre-bomb shelf water was -48 ±3 ‰ (triangle). The ice camp results are considered to be equivalent to Canada Basin water in that all are on the west side of Lomonosov Ridge. Our age model uses $\Delta R = 0 \pm 100$ (1σ) for the Holocene and Bolling/Allerod, within uncertainty of the pre-bomb estimate[30], but we use a larger ΔR for the YD (200±100) (Fig. ED5).

Fig. 5. Comparison of deglacial $\delta^{18}O$ between Orca Basin in the Gulf of Mexico and Beaufort Sea. Arctic data are based on *N. pachyderma* (s) (blue squares, core 15/27; red line, JPC-09) and Orca Basin data are based on the planktonic foraminifer

Globigerinoides ruber (green line, Williams et al.[5]; black squares, Leventer et al.[54]) (see ED for chronology details of the Leventer et al. core). The eastern Beaufort Sea freshened at about 12.9 ka coincident with the end of Gulf of Mexico freshening and consistent with the hypothesis that meltwater was diverted from the Gulf to a more northern outlet as deglaciation progressed[4]. YD= Younger Dryas, B/A = Bolling/Allerod, HS-1 = Heinrich Stadial 1, LGM = last glacial maximum.

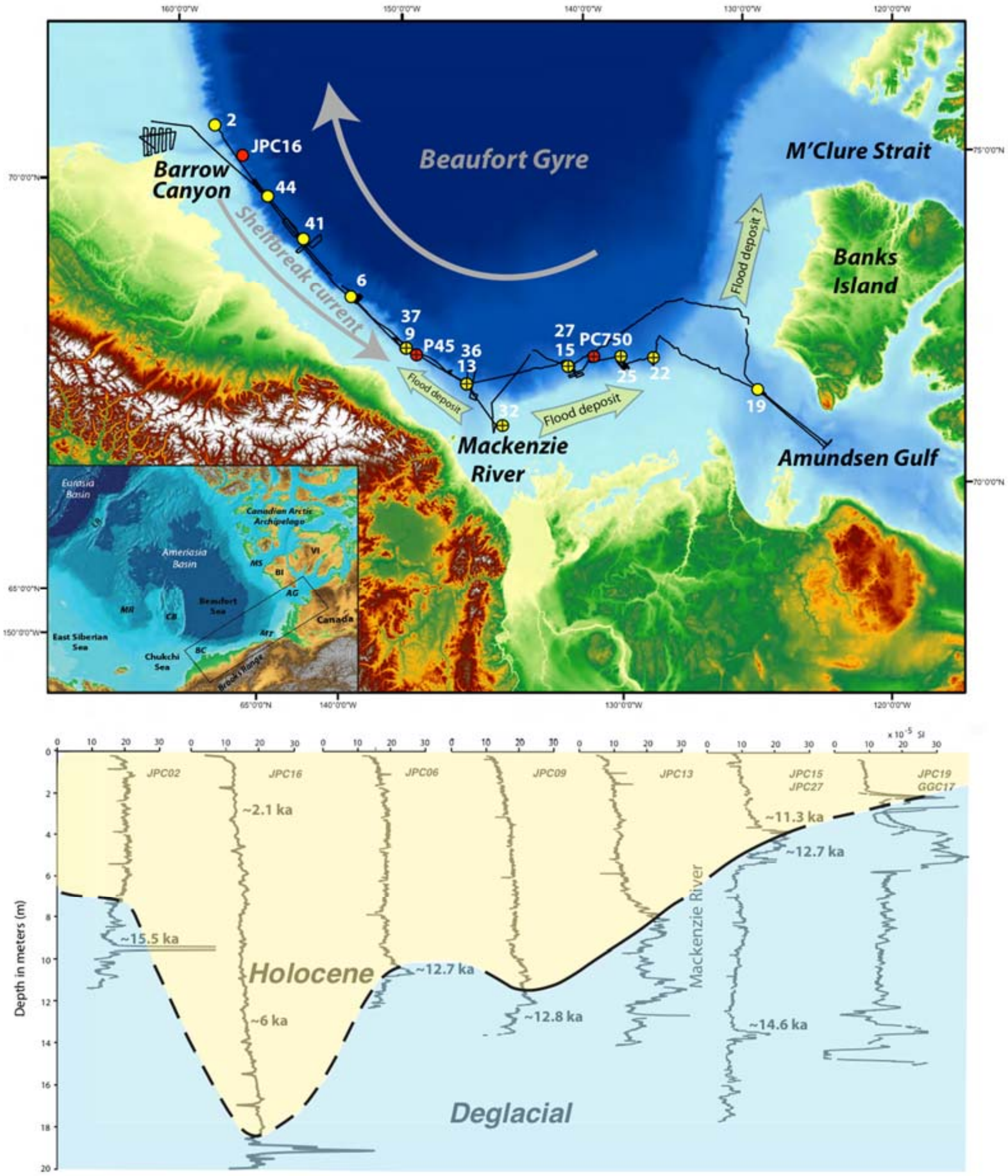
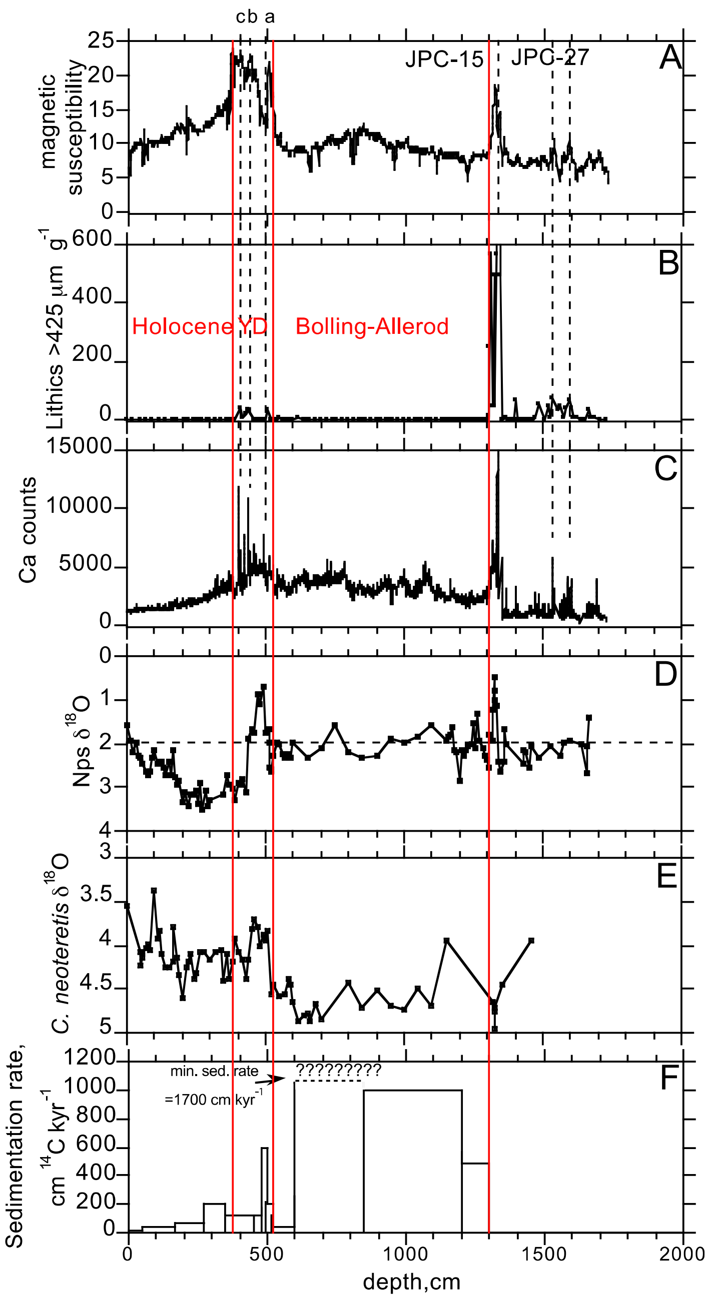

Beaufort Gyre
Barrow Canyon
Shelfbreak current
M'Clure Strait
Banks Island
Flood deposit ?
Flood deposit
Flood deposit
Mackenzie River
Amundsen Gulf
JPC16
P45
PC750
2
44
41
6
37
9
36
13
32
27
15
25
22
19
Eurasia Basin
Amerasia Basin
Canadian Arctic Archipelago
Beaufort Sea
East Siberian Sea
Chukchi Sea
Brooks Range
Canada
JPC02
JPC16
JPC06
JPC09
JPC13
JPC15
JPC27
JPC19
GGC17
~2.1 ka
~15.5 ka
Holocene
~6 ka
~12.7 ka
~12.8 ka
~11.3 ka
~12.7 ka
~14.6 ka
Mackenzie River
Deglacial
Depth in meters (m)
x 10^-5 SI

Figure 1



c b a
JPC-15
JPC-27
A
magnetic susceptibility
B
Lithics >425 µm g-1
Holocene
YD
Bolling-Allerod
C
Ca counts
D
Nps δ18O
E
C. neoteretis δ18O
F
Sedimentation rate, cm 14C kyr-1
min. sed. rate
=1700 cm kyr-1
??????????
depth,cm

Figure 2

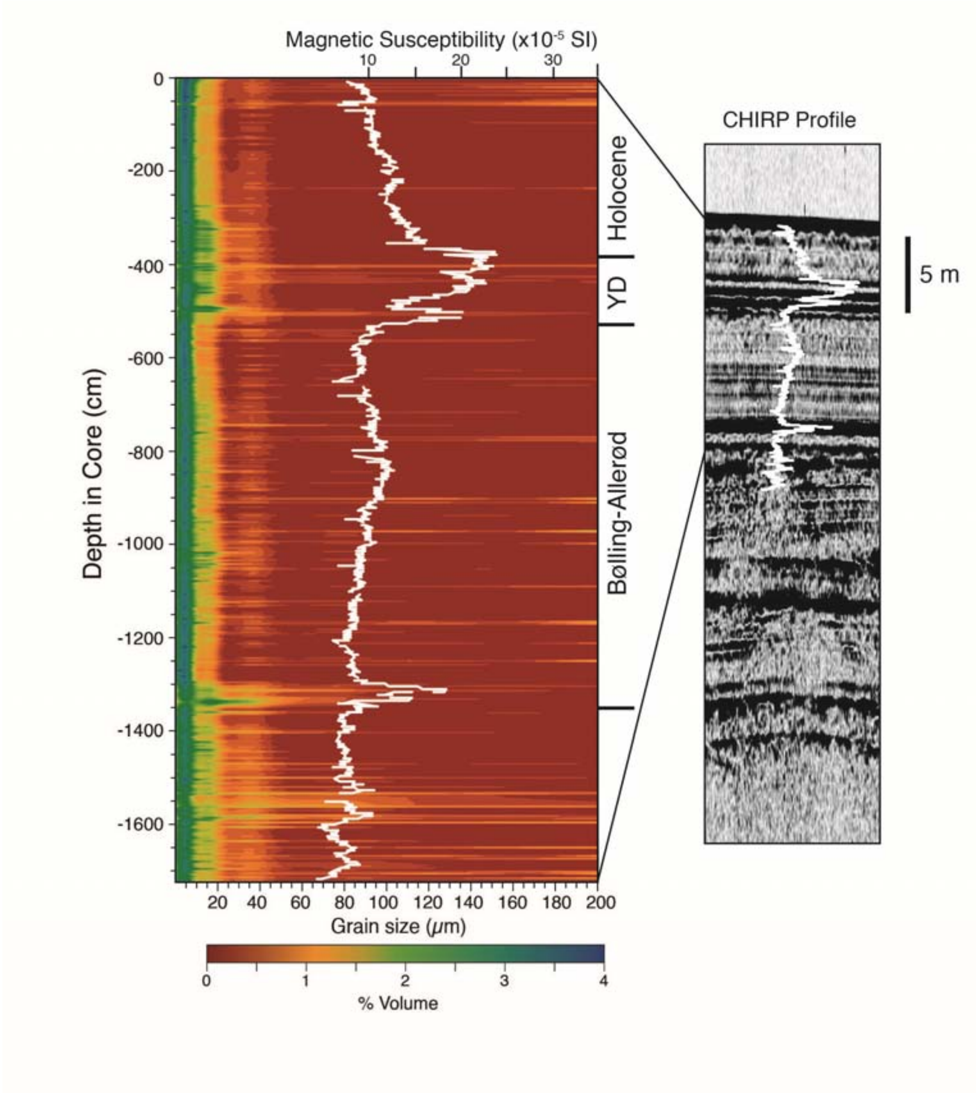

Magnetic Susceptibility (x10-5 SI)
10
20
30
0
-200
-400
-600
-800
-1000
-1200
-1400
-1600
Depth in Core (cm)
Holocene
YD
Bølling-Allerød
20
40
60
80
100
120
140
160
180
200
Grain size (μm)
0
1
2
3
4
% Volume
CHIRP Profile
5 m

Figure 3

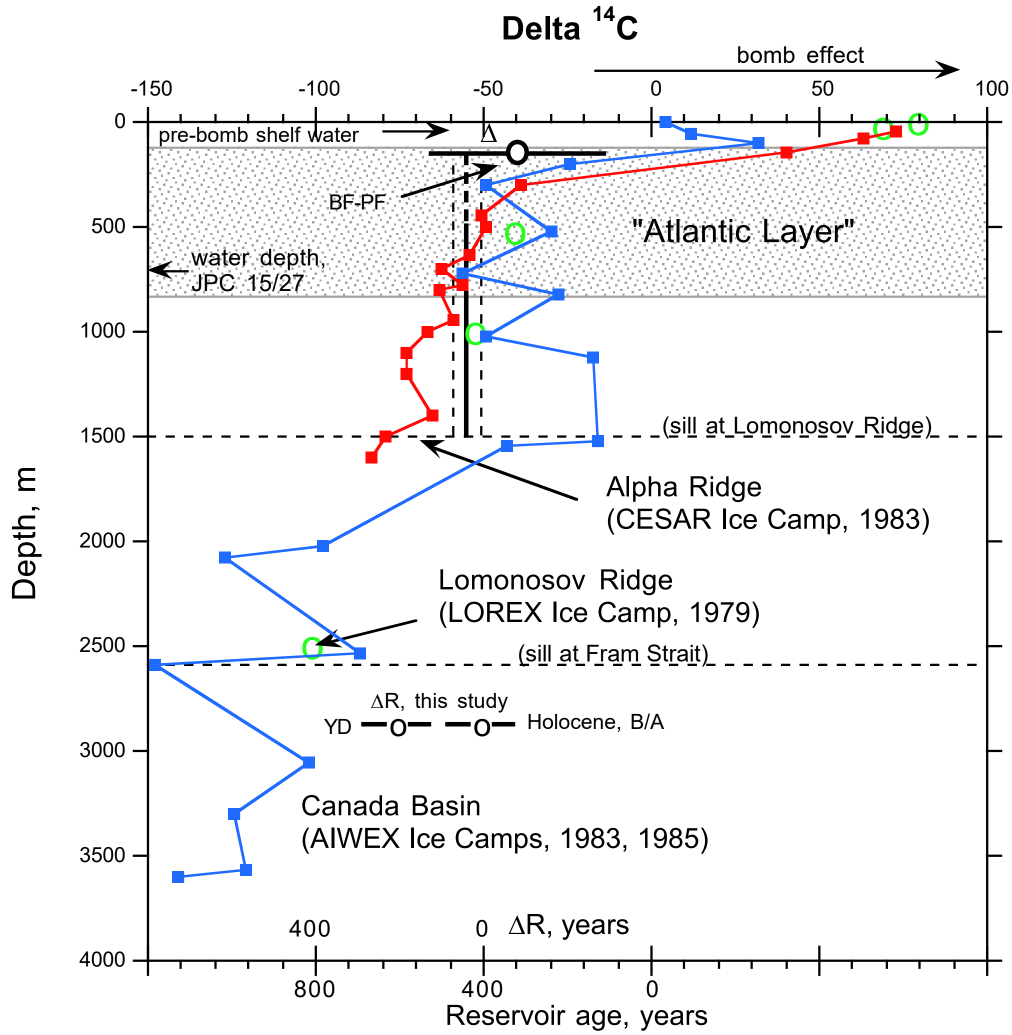

Delta 14C
bomb effect
-150
-100
-50
0
50
100
0
500
1000
1500
2000
2500
3000
3500
4000
pre-bomb shelf water
Δ
BF-PF
"Atlantic Layer"
water depth,
JPC 15/27
(sill at Lomonosov Ridge)
Alpha Ridge
(CESAR Ice Camp, 1983)
Depth, m
Lomonosov Ridge
(LOREX Ice Camp, 1979)
(sill at Fram Strait)
ΔR, this study
YD
Holocene, B/A
Canada Basin
(AIWEX Ice Camps, 1983, 1985)
400
0
ΔR, years
800
400
0
Reservoir age, years

Figure 4

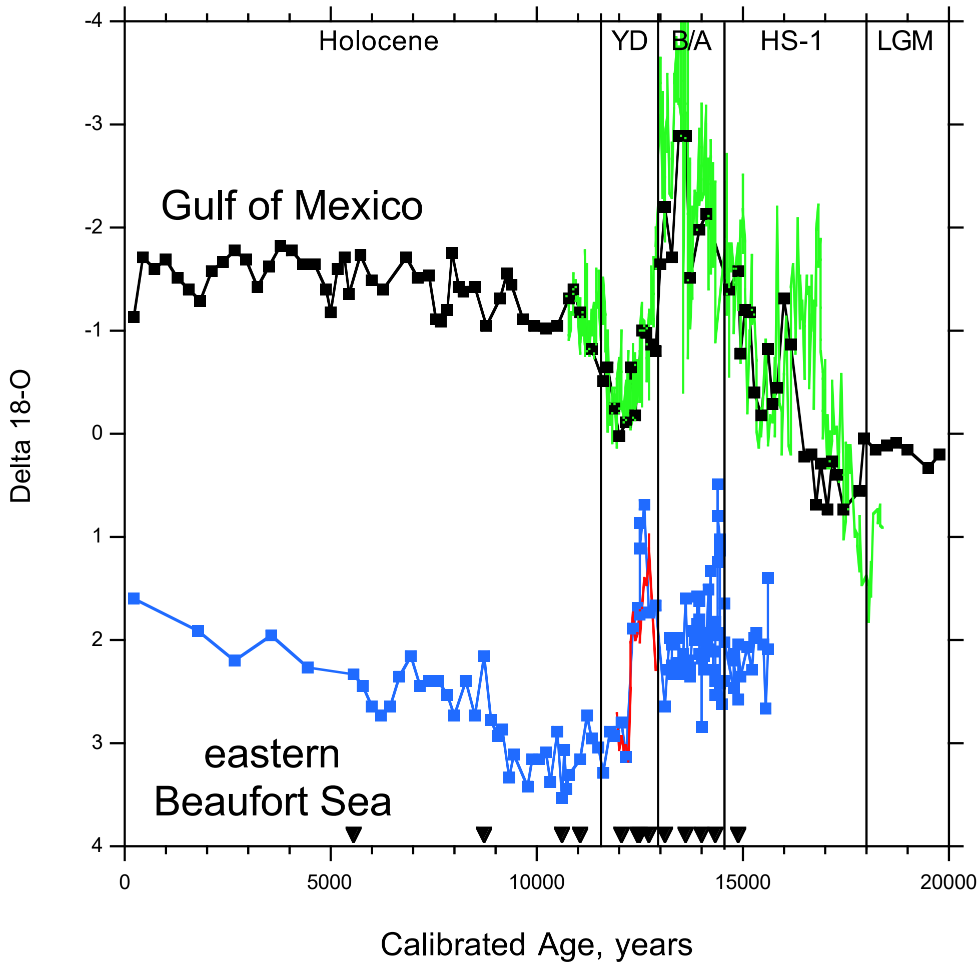


Figure 5

Deglacial Floods in the Beaufort Sea Preceded Younger Dryas Cooling

Supplementary Information

Extended Data

1. Site surveying

Sites were surveyed on USCGC Healy using the hull mounted multibeam swath bathymetry system, and a Knudsen 320B/R sonar. The Knudsen system, also hull mounted, operates at a central frequency of 3.5 kHz and sweeps between 2 and 6 kHz which makes it “chirp.”

2. Stratigraphy

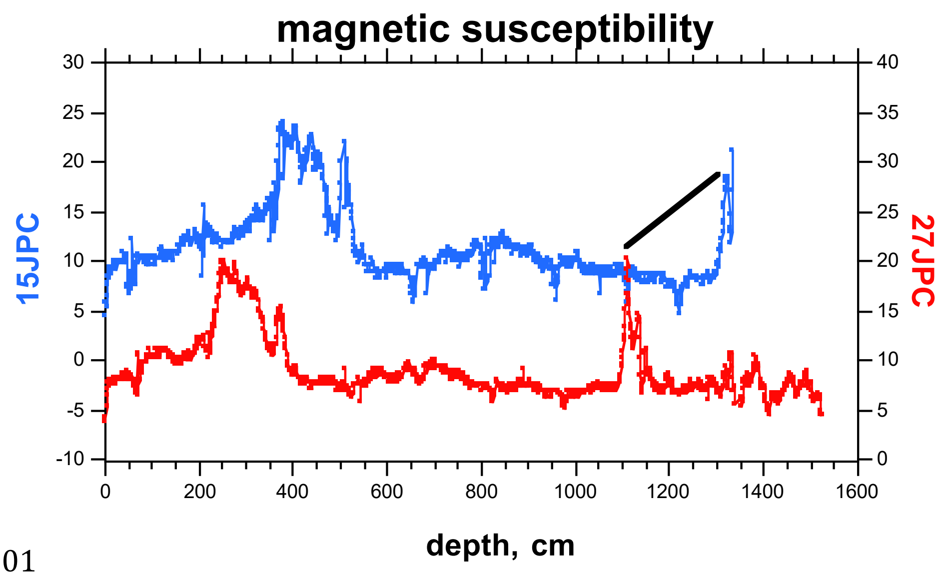


Fig. ED1. Magnetic susceptibility records of HLY1302 cores JPC15/27 from the same location at 690 m on the continental slope east of Mackenzie River (JPC15: 71°06.222'N, 135°08.129'W; JPC27: 71°06.360'N, 135°09.640'W). To make a 1729 cm composite section, we patched to JPC-15 at 1329 cm the data below 1125 cm in JPC-27 (with a +205 cm offset).

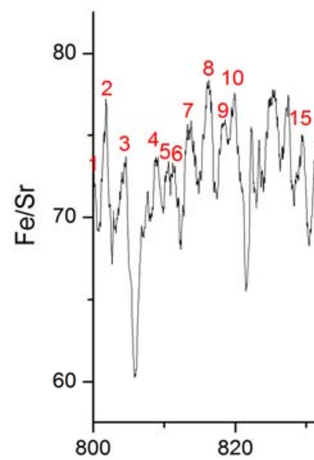


Fig. ED2. Laminae counted using Fe/Sr variability of a one-meter section in HLY1302 JPC15. Many other elemental pairs show similar variability. High Fe/Sr suggests greater terrestrial content. The resolution of the data is 0.4 mm and the data are smoothed with a 19-point running mean. There are about 50 peaks in this section with 2 cm/cycle on average, and the number of cycles varies little with counting method. We counted ~300 laminae between 600 cm (13460 ka) and 1201 cm (14408 ka) where the deposition rate is uniformly high, and those reflect ~300 oscillations in terrigenous input to the continental slope that are probably not annual (300 laminae/948 years = 0.32 laminae/yr) unless the age model underestimates the rate of sedimentation. Note that the calendar ages give lower accumulation rates than those using conventional ^{14}C years as in Fig. 2F.

3. Sampling and stable isotopes

Core JPC15 was initially chosen for study because of its position east of Mackenzie Trough and because of its typical looking magnetic susceptibility. Not knowing what was present, we began with samples ~20 g dry every 50 cm. Based on early $\delta^{18}O_{Nps}$ results, sampling was increased to every 10 cm. About 20 clean and clear (not infilled) specimens of Nps were chosen for stable isotope measurements using standard methods[55]. Although the focus of the stable isotopes in this paper is $\delta^{18}O$, $\delta^{13}C$ was measured and is reported in **Table ED2**. Note that the $\delta^{13}C$ data are featureless for both Nps and *C. neoteretis*. They compare well with the $\delta^{13}C$ of dissolved inorganic carbon reported from the eastern Beaufort Sea[56].

4. Chronology

4.1 Gulf of Mexico

Leventer et al.[54] was the first study to improve on the original Kennett and Shackleton[4] $\delta^{18}O$ data with a new higher resolution series from piston core EN32 PC6 in anoxic Orca Basin and with bulk organic ^{14}C dates. That was before AMS dating, so as the interest in meltwater diversion and the origin of the YD grew in the 1980s, Broecker et al.[9, 57] used AMS methods to redate the core. Unfortunately, their results contained substantial age reversals. We include Leventer et al.[54] data in Figure 5 because they provide a Holocene context for the higher resolution and better dated $\delta^{18}O$ series of Williams et al.[5]. The two data sets are in good agreement where they overlap. However, this was achieved by (selective) use of the available AMS dates at 29.5, 436.5, 471.0, 486.5, and 809 cm (refs 9, 57) and calibration using $\Delta R=0$.

4.2 Beaufort Sea

Levels for AMS dating (at NOSAMS) were identified based on the $\delta^{18}O_{Nps}$ results, and resampled so that as much as 80 g dry were picked to get sufficient Nps. Where possible, only clean specimens of Nps and *C. neoteretis* were selected from the size fraction >150 µm. This was easy for *C. neoteretis* because the test is transparent**,** but for Nps we set aside clean and empty specimens and cleaned the remainder mechanically as described elsewhere[58]. If that did not clean them sufficiently, we cleaned them ultrasonically, always setting aside clean ones at each step. Ultrasonic cleaning broke up most tests, but clean fragments were sometimes selected for inclusion in the dated sample. Our chronology is based on a Bayesian age model (Figure ED6) using the Nps dated levels (**Table ED1**, **Fig. ED3**).

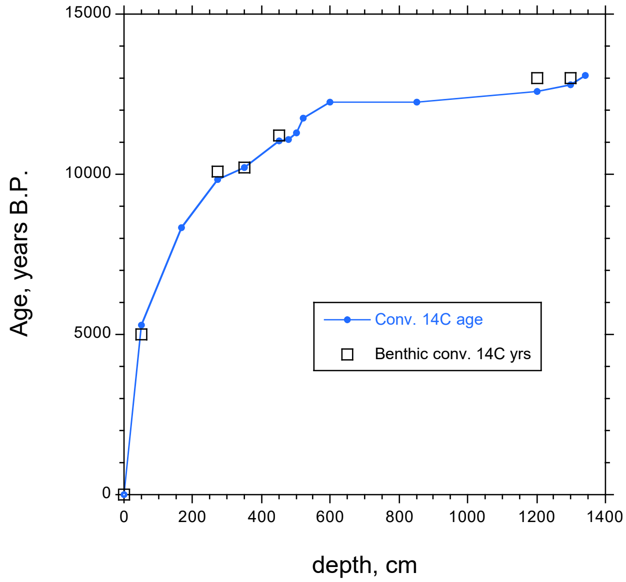


Fig. ED3. Age-depth relationship of conventional AMS ^{14}C dates on Nps (blue circles) and *C. neoteretis* (open black squares) from JPC-15.

4.3 Choice of ΔR

A. Pre-bomb ΔR

Any discussion of ΔR should begin with the modern, or better yet, the pre-bomb ocean. For the Beaufort Sea, the pre-bomb ΔR has been estimated based on radionuclide tracers for Arctic processes[30, 59] (Fig. 5), and in pre-bomb museum specimens of mollusks (especially bivalves[60]). These are very different data sets and the resulting ΔRs are not directly comparable because the mollusk data came from specimens collected along the nearshore continental shelf whereas the ice station data (and our core sites) are far offshore. One notable thing about the Ostlund et al.[30] analysis is discussion of the ^{14}C measurement on surface waters in the east Greenland Current in 1957 that leads them to "safely assume" that shelf water had a pre-bomb $\Delta^{14}C$ of -48 ± 3 ‰. (These data were published first by Fonselius and Ostlund[61] before international standardization.) Although east Greenland is about as far as you can get in the Arctic from the Beaufort Sea, Ostlund and Hut[59] showed that the residence time of shelf and near surface waters in the Arctic is only ~10 years. However, they had no shelf water data from the west Arctic where there

Fig. ED4. Locations of pre-bomb bivalve data[60] from off Alaska on left, downstream in the Amundsen Gulf (middle), and far to the east in Foxe Basin. These sites were chosen to define a flow path where Bering Strait water always hugs the coast and turns right. Today the shelfbreak current has been traced to the entrance of Amundsen Gulf[36], but the bivalve ^{14}C data have a Pacific signature as far to the east as northern Foxe Basin.

is low "preformed" $\Delta^{14}C$ from the Pacific, based on the bivalves.

McNeely et al.[60] compiled mollusk ^{14}C data from all around Canada for the specific purpose of knowing ΔR at continental shelf depths. In the Beaufort-Chukchi Seas they reported dates on 7 bivalve specimens collected from two stations (**Fig. ED4**). Six bivalves were suspension feeders and one was a deposit feeder; that one is significantly older than the others (ΔR=610 yrs). Excluding that datum, the others have a mean ΔR of 440±101 yrs, or a mean $\Delta^{14}C$ close to -100 ‰. That result is greatly different than the $\Delta^{14}C$ of -48 ‰ directly measured in in East Greenland shelf waters[30].

The missing element in the Ostlund and Hut[59] and Ostlund et al.[30] analysis was a source of relatively old waters from the NE Pacific via the Alaska Coastal Current and, through Bering Strait, to the shelf break current in the Beaufort Sea. The shelf break current can be traced as far east as Amundsen Gulf, by which point it is dissipated without evidence of entering the Gulf[36], but the pre-bomb mollusk data[60] can be used to trace transport to the Labrador Sea through the Canadian archipelago in recent times. Forty $\Delta^{14}C$ measurements of Pacific mollusks (Victoria, BC to Bering Strait), excluding deposit feeders, average ΔR=388 ±86 yrs, not significantly different from the Chukchi/Beaufort value cited above (440 ±101 yrs). By Amundsen Gulf, where

McNeely et al.[60] have 7 observations from 5 sites (Fig. ED4), the result, ΔR=350±116, is within uncertainty of the Bering Strait source waters. However, by Foxe Basin, ΔR=286 ±74 yrs (n=8), significantly lower (younger) than the Beaufort/Bering Strait data. We choose Foxe Basin as an end point because it represents a pathway that is least likely to encounter younger Atlantic shelf waters, and for the same reason we only use those data on the south side of the strait that connects Gulf of Boothia to Foxe Basin. Nevertheless, a trend of increasing $\Delta^{14}C$ in pre-bomb mollusks from the Gulf of Alaska to Foxe Basin suggests mixing with a young North Atlantic component. These data are substantially older than the East Greenland mollusks, where ΔR=92 ±67 years (n=12).

The east Greenland shelf is the only place where pre-bomb $\Delta^{14}C$ has been measured in both shelf waters (-48±3 ‰) and in mollusks (-61±7 ‰), and with results in reasonable agreement. However, this does not mean that shelf ΔR should be used to calibrate ^{14}C ages from foraminifera on the Beaufort continental slope for a few reasons. (1) The shelfbreak waters that carry the old signal from Bering Strait are well inshore of the surface water overlying our core sites[36]. (2) Although we do not know the pre-bomb ^{14}C age of Beaufort Sea surface waters (Fig. 4), the rather close agreement of paired benthic and planktonic ^{14}C ages suggests the planktonics live in water influenced by the Atlantic layer even in the Holocene. During pre-Holocene time (>11 ka), before Bering Strait was flooded[26, 27], the Atlantic layer might have shoaled in the absence of Pacific water, all else being equal (pers. comm. 2016 from R. Pickart and M. Spall). However, most importantly, (3) the absence of old Pacific water in the pre-Holocene Arctic means that shelf waters must have had much lower ΔR than today prior to 11 or 12 ka.

B. **ΔR in the Nordic Seas**

There are no data from the western Arctic Ocean that can be used to estimate ΔR, so we turn to the Nordic Seas where waters feeding the Arctic surface circulation flow northward along the coast of Norway, and southward from Fram Strait along the Greenland coast to the Labrador Sea. The only useful data sets for this come from Bondevik et al.[28] and Cao et al.[29]. Cao et al.[29] synthesized existing ^{14}C data from the high latitude North Atlantic and presented new data on solitary corals from Orphan Knoll (1600 m water depth). They concluded that the Allerod warm period had a ΔR similar to today (~0 years), and that ΔR was likely about 200 yrs greater during the YD. The Bondevik et al.[28] data contributed greatly to that conclusion. Orphan Knoll data do not reflect coastal conditions but rather the ventilation of the central Labrador Sea, with an unknown transit time from the surface to ~1600 m.

Accordingly, we base our surface reservoir corrections for the eastern Beaufort Sea on the Bondevik et al.[28] data for samples where pairs of marine and terrestrial (atmospheric) ^{14}C dates came from within 1-cm of each other in their cores, and not including data the authors rejected as coming from out-of-place fossils. The difference in conventional ^{14}C age between the marine and terrestrial data is defined as ΔR, and terrestrial dates have been recalibrated using Calib 7.1 using the Marine 2013 curve.

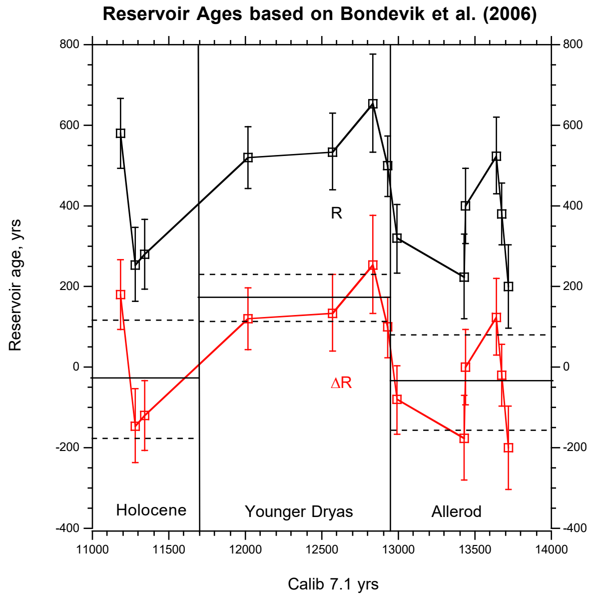


Fig. ED5. Summary of R and ΔR for the Allerod through the early Holocene based on pairs of marine and terrestrial ^{14}C dates from Bondevik et al.[28]. Vertical error bars are ±1σ. Horizontal lines show mean values for the three time intervals, with dashed lines representing ±1σ.

For the Allerod there are seven marine-terrestrial pairs of AMS dates[28] that return a mean of -36 ± 116 years (1σ) (**Figure ED5**). For the Younger Dryas and Holocene, the statistics are 170 ± 60 (n=3) and -28 ±148 (n=3), respectively. Although these data essentially come from only one location and are highly variable, they are the only dates that meet our requirement of being in the flow of coastal waters either entering or leaving the Nordic Seas. The low Allerod and higher YD ΔR are consistent with the synthesis of Cao et al.[29] and it makes sense that, during that relatively warm period with better North Atlantic ventilation, the reservoir effect would have been similar to the Holocene. The Holocene results are generally concordant with the pre-bomb estimate of Ostlund[30]. For calibration purposes we chose ΔR = 0 ± 100 years for the Holocene and the Allerod, and 200 ± 100 for the YD. These values are increased somewhat from the measured values (Fig. ED5) because there is some evidence for increased ΔR with latitude along the Norwegian coast, but even the authors who made that observation do not agree about its significance[62]. Note that for dating the beginning of the YD it is important to use the Allerod ΔR, not that of the YD. This is because if the YD flood caused a decrease in the AMOC, and if that caused the increase in ΔR through changes in storage and exchange in the ocean-atmosphere carbon system, then the Allerod ΔR is more appropriate than the YD ΔR.

4.4 Bayesian age modeling.

As recommended by an anonymous reviewer, we developed an age model for JPC15/27 using the "Bacon" software of Blaauw and Christen[63] (2011). This method evaluates rates of sedimentation for discrete sections of the core, and these are informed by results in surrounding sections. The appropriate command settings for our model are: Bacon("JPC-15", 25, acc.mean=2, acc.shape=1.1, normal=TRUE, remember=FALSE, depths.file=T), agedepth(rotate.axes=TRUE, rev.yr=TRUE). We input our ^{14}C dates with the higher ΔR during the YD, we fixed the core top to equal zero years, and the calibration was done using the Marine 13 curve. The resulting age-depth relationship (**Fig. ED6**), illustrates the mean age of levels in the core and the 95% confidence interval.

Of critical importance is the calendar age of the sample at 514 cm, where $\delta^{18}O$ is about halfway to its minimum value: 12,939 calendar years B.P., with a minimum age of 12,786 years and a maximum age of 13,080 years, or about 12.94 ± 0.15 ka. The abrupt decrease in $\delta^{18}O$ lies within the 95% confidence interval of 13 ka, the nominal date for the diversion of meltwater from the Gulf of Mexico[5], and before the ~12,850 year start of the YD on the Greenland ice core timescale[34].

We experimented with other age models, to test the robustness of our result. Using a constant ΔR of 0 ± 100 yrs for the entire record gave about the same age for the sample at 514 cm with the variable ΔR model, so we know that the "bacon" age is not influenced by the decrease of sedimentation rate and increase in ΔR during the YD. Likewise, doubling the uncertainty in ΔR for the entire record returns the same ages but with less confidence. In sum, our conclusions are driven mostly by the choice of ΔR; we cannot reject the hypothesis that the flood down Mackenzie River was coincident with the beginning of the Younger Dryas cooling using any DR that is consistent with the Allerod data (Fig. ED5). Using our preferred age model (Fig. ED6), we summarize ages and uncertainties associated with the $\delta^{18}O_{Nps}$ evidence for the YD flood in JPC15/27 in **Table ED3**.

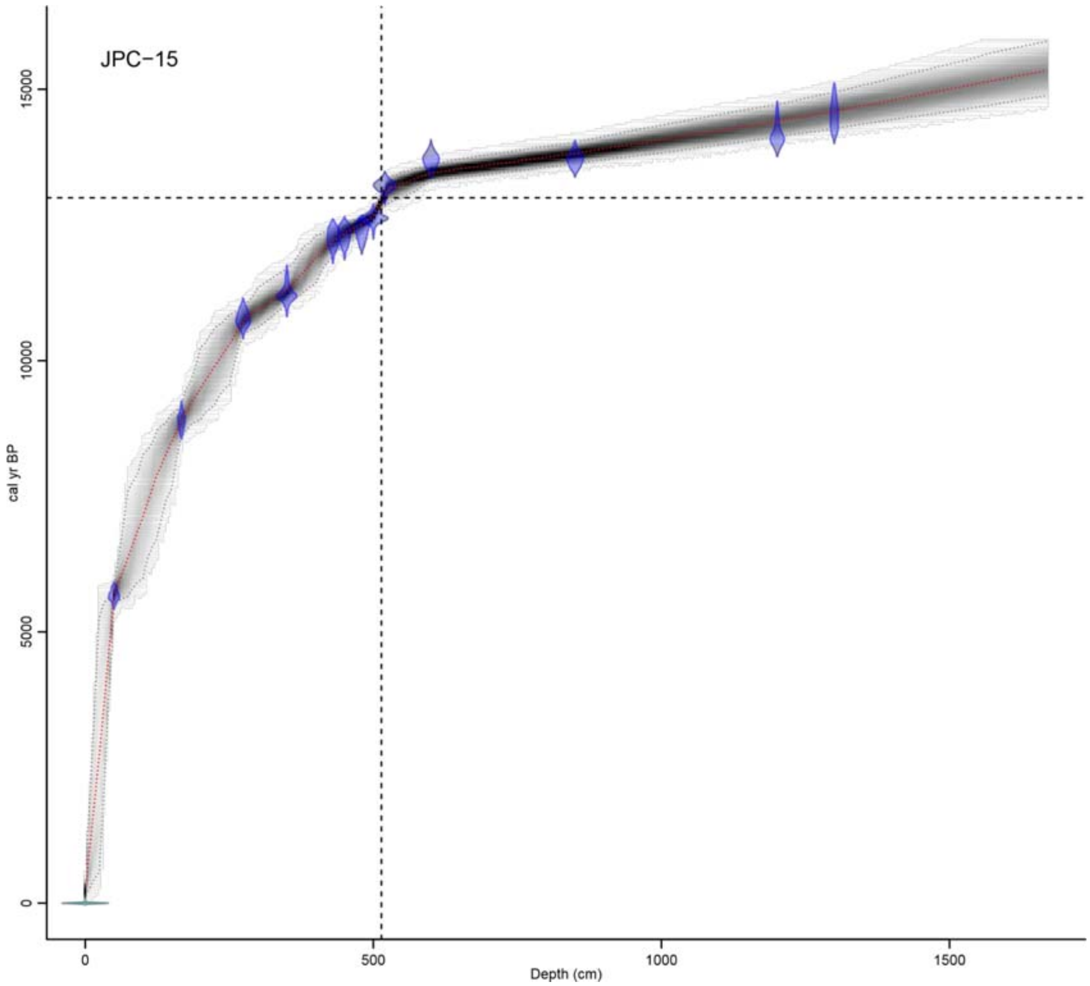
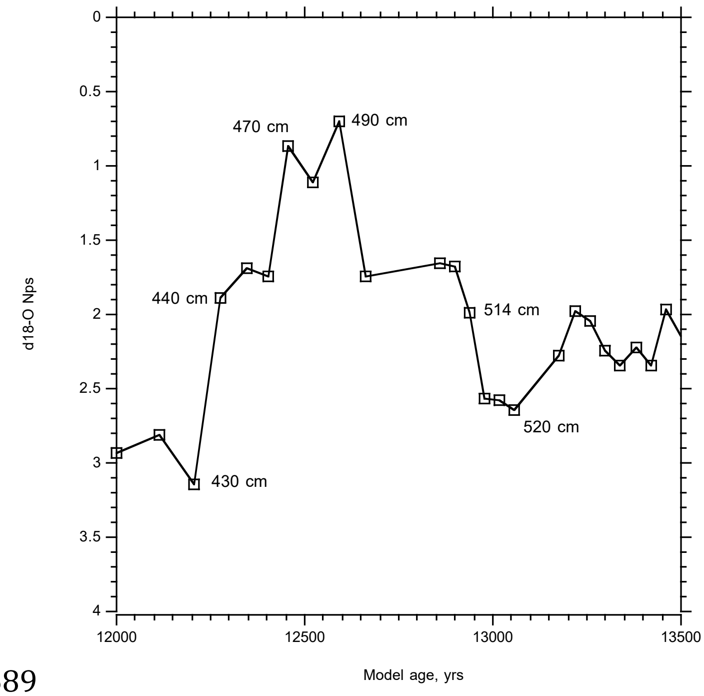


Figure ED6. Age model for JPC15/27 using the Bayesian method “Bacon”[63]. Horizontal dashed line is at 13 ka and vertical dashed line is at 514 cm. The model gives an age at 514 cm of 12.94 ± 0.15 ka. A blow-up of the critical $\delta^{18}O_{Nps}$ data 12-13.5 ka is shown in Figure ED7.

5. Regional summary of oxygen isotope data

5.1 New core data from this study

It is important to determine the spatial extent of the YD flood within the Beaufort Sea because Coriolis forcing would drive a buoyant flow to the right from Mackenzie River, and northward along the Canadian Archipelago toward Fram Strait. Such a direct path to the North Atlantic might have the most climate impact because the surface waters would be freshest. On the other hand, wind forcing could counteract the Coriolis driven



ED Fig. 7. Blow-up of $\delta^{18}O_{Nps}$ data associated with the YD flood at JPC15/27. The depth of important features is indicated for reference to ED Table 3. We interpret these data to mean that the flood was underway as early as 12940 ± 150 years ago, the age of the sample at 514 cm.

flow and perhaps allow more mixing with Beaufort Gyre. In that case, the freshening in the North Atlantic region might have been less but may have lasted longer.

Here we summarize the stratigraphic data from cores extending from JPC15/27 in the east, which we consider to be a "type section," to other cores as far west as Barrow, AK (**Fig. ED8**). West of Mackenzie River at JPC-09 we have identified a $\delta^{18}O_{Nps}$ minimum at about 13 m below the seafloor. It reaches 1 ‰, close to the minimum at JPC15/27 and it occurs a meter below a prominent maximum in magnetic susceptibility. This phasing is similar to results at JPC15/27, and the AMS date at JPC-09 falls within the range of dates constraining the flood event to the east. The brief peak in magnetic susceptibility at JPC15/27 at ~500 cm is not matched at JPC-09 probably because the ice rafting, which becomes common >1300 cm, stopped the corer. If we calibrate the YD ^{14}C ages from JPC-09 (Table ED1) with ΔR=200±100, the $\delta^{18}O_{Nps}$ changes are well-matched at the two cores (Fig. 5, Fig. **ED9**).

JPC-09 is very close to core P45 of Andrews and Dunhill (2004)(Fig. 1), so we recalibrated the age model for that core using ΔR= 0±100 (post YD) and plotted their $\delta^{18}O_{Nps}$ with the new data from this study (Fig. ED9). The agreement between these cores is good, although the age model may make the bottom of P45 too old because their oldest date was on benthic foraminifera. Note that the minimum in $\delta^{18}O_{Nps}$ was not found by Andrews and Dunhill (2004), most likely because the corer failed to penetrate the ice rafted layer at about 5 m subbottom.

Continuing farther west of Mackenzie River, the $\delta^{18}O_{Nps}$ at JPC-06 records only a small minimum before the main peak in magnetic susceptibility (Fig. ED8). This suggests that the YD meltwater plume must have been very localized to the region east of this site with only minor salinity lowering of the near surface ocean. Of the samples examined, a small peak in ice rafting is associated with the small minimum in $\delta^{18}O_{Nps}$.

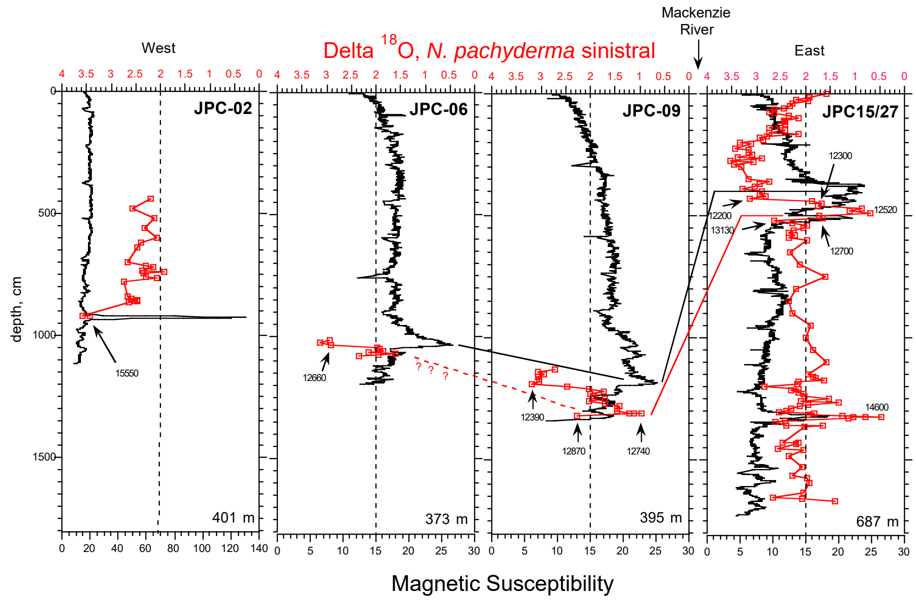


Figure ED8. Comparison of magnetic susceptibility and $\delta^{18}O_{Nps}$ stratigraphies in a zonal transect of cores from east of Mackenzie River (JPC15/27) to Barrow, AK (JPC-02). The vertical dashed line in each core marks the baseline $\delta^{18}O_{Nps}$ as reference for surface freshening. The solid black and red lines correlate the magnetic susceptibility $\delta^{18}O_{Nps}$ peaks, respectively. Note the decreasing influence of $\delta^{18}O_{Nps}$ lowering (freshening) from east to west.

In the Chukchi Sea off Barrow, AK, the most notable feature of JPC-02 is an IRD and magnetic susceptibility peak at ~920 cm that dates to 15.55 calibrated ka and includes a 6-cm dark non-carbonate dropstone. Because this event is not recorded far to the east at JPC15/27, and is >1000 years older than the 14.6 ka event at that site, it gives a maximum age for the bottom of the composite section at JPC15/27, assuming the event came from the Canadian Archipelago and would probably have spread across the Beaufort Sea. That maximum age (15.5 ka) agrees well with the 15.4 ka extrapolated age for the end of the JPC15/27 $\delta^{18}O_{Nps}$. Also of note is the maximum in $\delta^{18}O_{Nps}$ coincident with this IRD layer; this is the opposite of what we see in the YD and 14.6 ka events closer to Mackenzie River and it is the heaviest we have measured in this study.

Most of the $\delta^{18}O_{Nps}$ data fall higher than the 2 ‰ reference level for the entire record <15.5 ka (Fig. ED8), similar to the nearby Holocene results from Keigwin et al.[26]. Thus, taking into account the ice volume effect on $\delta^{18}O$, we conclude that the near sea surface off Barrow was fresher than today during most of the deglaciation, but there must also have been a salinity gradient from the Chukchi Sea to the eastern Beaufort Sea. This points to Mackenzie River as the source of the freshening, but the absence of evidence for the YD flood off Barrow suggests that floodwaters were not diluted much by mixing in the Beaufort Gyre. If supported by further data, this could mean that the YD flood was brief compared to the mixing time of the Beaufort Gyre and might have been especially potent in affecting the AMOC.

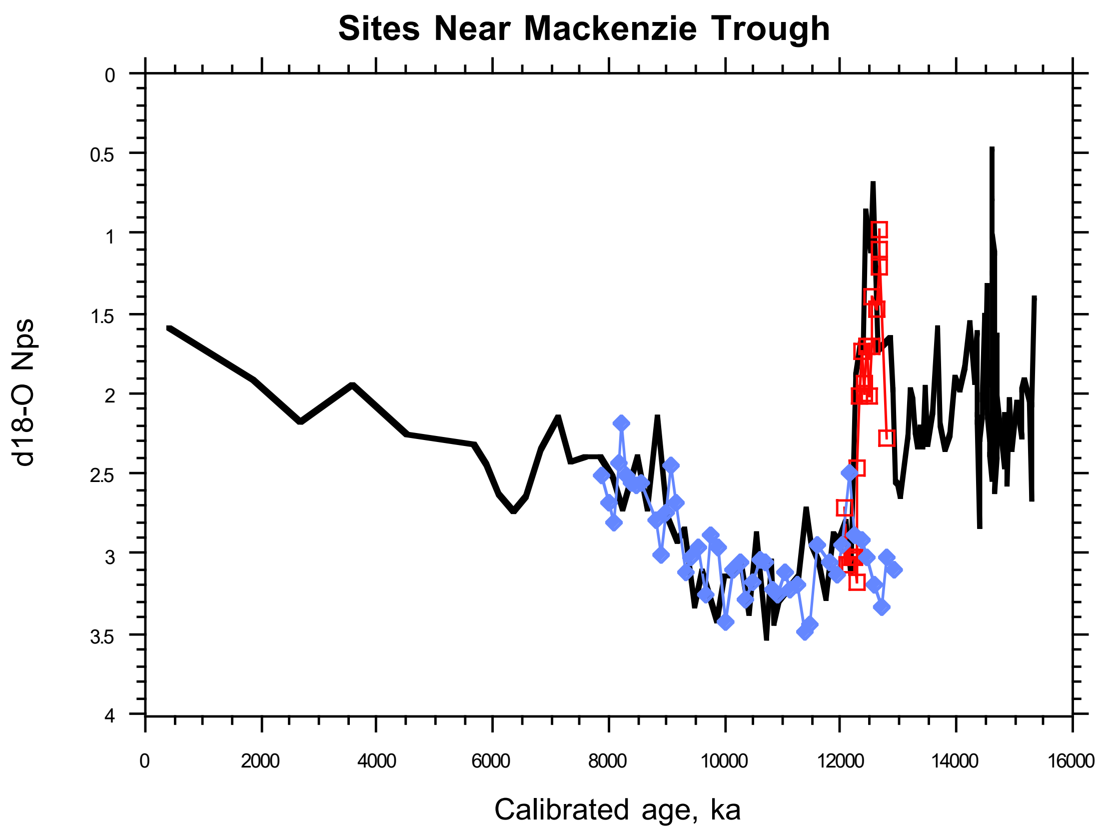


Fig. ED9. Comparison of $\delta^{18}O_{Nps}$ data between JPC15/27 (black line), JPC-09 (red squares) and core P45 (blue diamonds)[64]. Note the excellent agreement of minima in $\delta^{18}O_{Nps}$ from this study.

5.2 Other published core data

Several papers report stable isotope and radiocarbon data from the western Arctic (Mendeleyev Ridge) including, for example, Poore et al.[65]and Polyak et al.[66]. We cannot directly correlate our results from the eastern Beaufort Sea with those because they have much lower rates of sedimentation and fewer ^{14}C dates. Given that we also cannot correlate to our own core off Barrow (**Fig. ED8**), which does have high rates, it is possible that there was substantial spatial variability in near surface ocean conditions in the western Arctic during deglaciation. As an example of this, both Poore et al.[65] and Polyak et al.[66] found deglacial minima in $\delta^{18}O_{Nps}$ that are 0 ‰ or even lower. These are probably not evidence of the YD flood from Mackenzie River because the $\delta^{18}O_{Nps}$ is lower than we observe closer to the source, and the rates of sedimentation are probably too low to resolve such a brief event.

Closer to the Beaufort Sea, on the Chukchi Borderlands, Polyak et al.[67] do find a $\delta^{18}O_{Nps}$ minimum of about 1‰ that could be related to one of those we see at core 15/27.

However, using ΔR=0, their benthic foram calibrated date for that event is 13.8 ka which falls between the events we have found. That event is associated with a small peak in ice rafting (but not magnetic susceptibility), and below that there is a much larger undated IRD event coincident with a large peak in magnetic susceptibility.

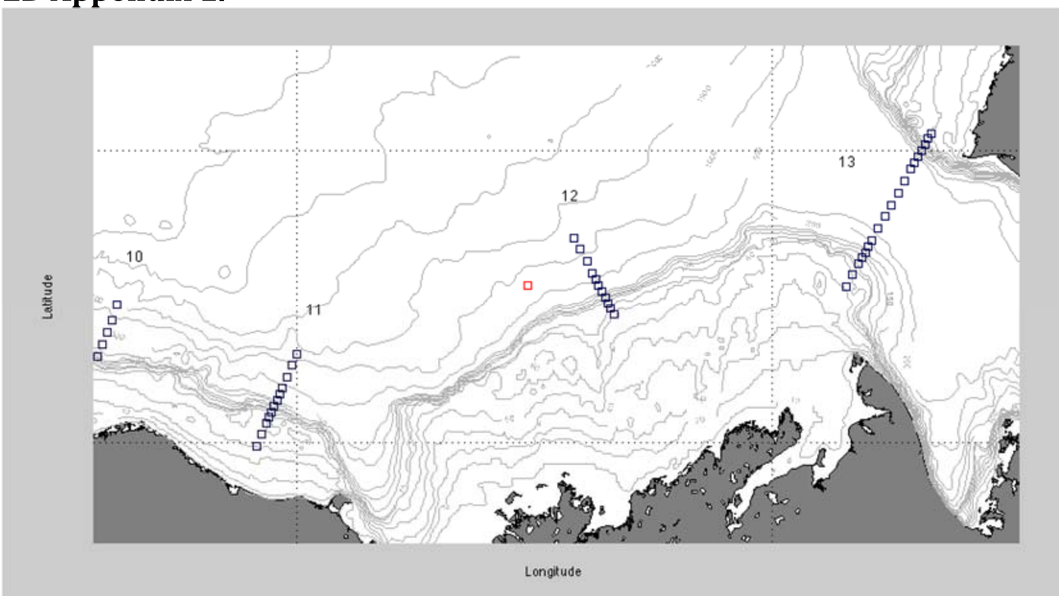
In addition to the comparisons discussed above, we can also correlate to results from Mackenzie Trough near our JPC-13 (ref. 41). One of their cores sampled the same high $\delta^{18}O_{Nps}$ (3.11 ±0.28 ‰, n=9) interval ~10-12 ka as in JPC15/27. The Schell et al. [41] data fall mostly between 10.9 and 10.6 ka when recalibrated.

Extended Data references:

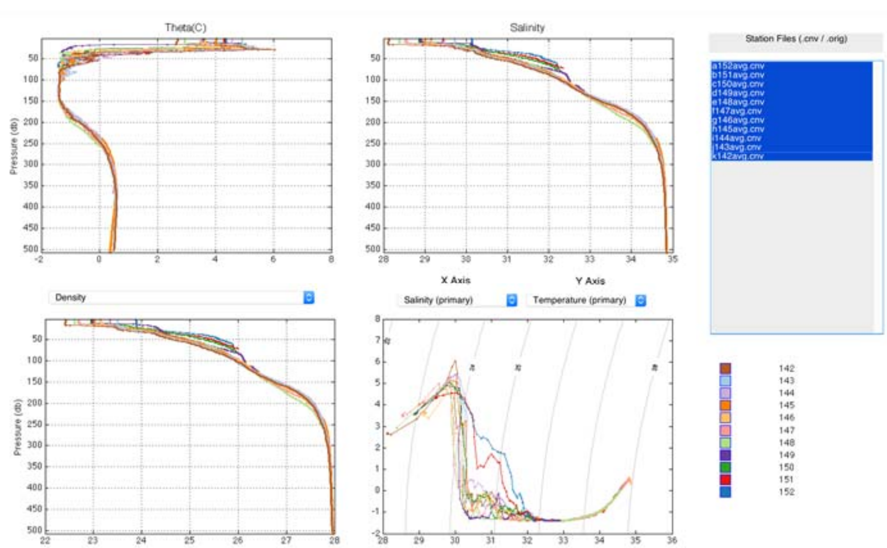
55. Keigwin, L.D. Radiocarbon and stable isotope constraints on last glacial maximum and Younger Dryas ventilation in th western North Atlantic. *Paleoceanography* **19**, doi: 10.1029/2004PA001029 (2004).
56. Oxtoby, L.E., Mathis, J.T., Juranek, L. W. & Wooller, M. .J. Estimating stable carbon isotope values of microphytobenthos in the Arctic for application to food web studies. *Polar Biol.* **39**, 473-483 (2016).
57. Broecker, W.S. et al. Acclerator mass spectrometric radiocarbon measurments on foraminifera shells from deep sea cores. *Radiocarbon* **32**,119-133 (1990).
58. Keigwin, L.D. & Lehman, S.J. Radiocarbon evidence for a benthic front near 3.1 km in the Equatorial Pacific Ocean. *Earth Planet. Sci. Lett.* doi: /10.1016/j.epsl.2015.05.025 (2015).
59. Ostlund, H. G. & Hut, G. Arctic Ocean water mass balance from isotope data. *Jour. Geophys. Research* **89**, 6373-6381 (1984).
60. McNeely, R., Dyke, A. S. & Southon, J. Canadian marine reservoir ages preliminary data assessment. *Geological Survey of Canada open file 5049*: DOI: 10.13140/13142.13141.11461.16649 (2006).
61. Fonselius, S. & Ostlund, G. Natural radiocarbon measurements on surface water from the North Atlantic and the Arctic Sea. *Tellus* **11**, 77-82 (1959).
62. Mangerud, J. Bondevik, S., Gulliksen, S., Hufthammer, A.K. & Hoisaeter, T. Marine ^{14}C reservoir ages for 19th century whales and molluscs from the North Atlantic. *Quaternary Science Reviews* **25**, 3228-3245 (2006).
63. Blaauw, M, & Christen, J. A. Flexible paleoclimate age-depth models using an autoregressive gamma process *Bayesian Analysis* **6**, 457-474 (2011).
64. Andrews, J. T. & Dunhill, G. Early to mid-Holocene Atlantic water influx and deglacial meltwater events, Beaufort Sea slope, Arctic Ocean. *Quaternary Res.* **61,** 14-21 (2004).
65. Poore, R. Z., Osterman, L., Curry, W. B. & Phillips, R. L. Late Pleistocene and Holocene meltwater events in the western Arctic Ocean. *Geology* **27**, 759-762 (1999).
66. Polyak, L., Curry, W. B., Darby, D.A., Bischof, J. & Cronin, T. M. Contrasting glacial/interglacial regimes in the western Arctic Ocean as exemplified by a sedimentary record from the Mendeleev Ridge. *Palaeogeo., Palaeoclim., Palaeoeco.* **203**, 73-93 (2004).

67. Polyak, L., Darby, D. A., Bischof, J. F., & Jakobsson, M. Stratigraphic contstraints on late Pleistocene glacial erosion and deglaciatiion of the Chukchi margin, Arctic Ocean. *Quat. Res.* **67**, 234-245 (2007).

ED Appendix 1.



ED Appendix Fig. 1. HLY1003 Station positions for the hydrographic data in ED Appendix Fig. 2 provided by Dr. Robert Pickart (WHOI). The position of HLY core JPC15/27 is shown as a red square.



ED Appendix Fig. 2. Data for section 12 presented vs pressure (~depth) as potential temperature (upper left), salinity (upper right), density (lower left), and T vs S (lower right). All hydrographic data are available at: http://aon.whoi.edu.