Dating the opening of the St. Lawrence Valley as a conduit for glacial meltwater and potential trigger of the Younger Dryas

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ABSTRACT

The trigger for the Younger Dryas climate reversal, the 1,200-year cold period that began 12.9 ka BP during the last deglaciation, is an unsolved problem in paleoclimate. A commonly proposed hypothesis for deglacial cold reversals holds that freshwater discharge from proglacial lakes into the North Atlantic slowed the Atlantic meridional overturning circulation (AMOC) and impeded surface ocean heat transport from lower latitudes. The path and volume of the discharge from the proglacial lakes remain unresolved, primarily between an eastward meltwater route into the Gulf of St. Lawrence and a northward route into the Beaufort Sea. We use acoustic surveys and radiocarbon-dated sediment cores from Lake Carmi and Fairfield Pond in the Champlain Valley in northern Vermont to test the viability of the eastward meltwater event as a trigger for the Younger Dryas. By dating the termination of proglacial lake deposition, we tightly constrain the retreat of the Laurentide ice sheet, and thereby provide a date for the opening of the St. Lawrence Valley. We use further radiocarbon dates and stratigraphic analysis to elaborate the deglacial chronology of the Champlain Valley. This study will establish the timing of freshwater access to the North Atlantic and, together with future work that will quantify the impact of freshwater on AMOC, permit us to evaluate the association between eastward melt and the onset of the Younger Dryas.

INTRODUCTION

The Younger Dryas is the canonical case of abrupt climate change. This cold episode interrupting deglacial warming is most visible in the Greenland ice cores; the isotope record from NGRIP, which has the highest resolution of the Greenland cores, indicates a 1,200 year-long cold period starting 12,900 calendar years BP and ending 11,700 cal years BP (Rasmussen et al 2006). Younger Dryas-age abrupt temperature changes appear in proxies spanning the globe (Shakun and Carlson, 2010) but the most dramatic changes occurred in the North Atlantic region. According to the NGRIP record, the initial cooling of 10˚C (to about 15˚C colder than modern temperatures) took place over decades and the warming of 15˚C at the end of the cold period occurred over decades.

This cooling has been interpreted as the product of a multitude of factors, which sort into three categories: ocean circulation, atmospheric circulation, and solar radiation (Fiedel, 2011). Most hypothesis involving oceanic and atmospheric reorganization involve the retreat of the North American and Fennoscandian ice sheets and the discharge of meltwater from the resulting proglacial lakes. Broecker et al. (1989) established reorganization of ocean circulation as the paradigm for the YD, proposing that discharge from glacial Lake Agassiz through the St. Lawrence Valley created a freshwater 'lid' on the North Atlantic. This layer of freshwater would have prevented deepwater formation, inhibiting Atlantic Meridional Overturning Circulation and thus the surface ocean transport of heat from lower latitudes. McManus et al. (2004) cite an excursion in the $^{231}Pa/^{230}Th$ ratio at YD onset as evidence of a sharp decline in meridional overturning.

Evidence casting doubt on the viability of an eastward drainage event—an absence of coarse flood deposits along the eastward route, evidence that the eastern outlets of Lake Agassiz were still glaciated at the start of the Younger Dryas (Teller et al., 2005), a muted freshwater signal in the isotope record from the Gulf of St. Lawrence (de Vernal et al., 1996)—has led some to focus on northward drainage from Lake Agassiz into the Arctic Ocean. A commonly proposed path for northward drainage involves the Mackenzie Valley, which discharges into the Arctic Ocean via the Beaufort Sea (Tarasov and Peltier, 2005). Condron and Winsor (2012) use an ocean sea-ice model to show that freshwater from the St. Lawrence Valley weakens AMOC by <15% while meltwater input into the Beaufort Sea weakens AMOC by >30%. Finally, Keigwin et al. (2018) identify a negative isotope excursion in Beaufort Sea sediments dated to 12.94 ka as evidence of substantial sea surface freshening.

As the body of evidence supporting northward meltwater as the YD trigger grows, the role of the Saint Lawrence Valley remains an open question. Key to developing a more concrete understanding of the contributions of eastward meltwater is establishing an age of the opening of the Saint Lawrence Valley as a conduit for glacial meltwater. As the Laurentide Ice Sheet (LIS) retreated downslope into the isostatically-depressed continent after the Last Glacial Maximum, glacial meltwater formed a series of proglacial lakes along the margin of the ice sheet. Glacial Lake Agassiz receives most attention as the source of meltwater, but there may have been enough freshwater stored in the Champlain lowlands to affect ocean circulation. Reconstructed paleoshorelines indicate decreasing lake levels in the lowlands over time, as meltwater breached a series of dams and expanded to fill the Champlain lowlands. Glacial Lake Candona, which formed when the Fort Ann sill breached approximately 13,000 calendar years ago BP, was the final lacustrine stage, dammed in the east by a southeastern lip of the LIS. When the retreat of the LIS finally permitted the draining of Lake Candona, marine water from the Gulf of St. Lawrence filled the lowlands and formed the Champlain Sea.

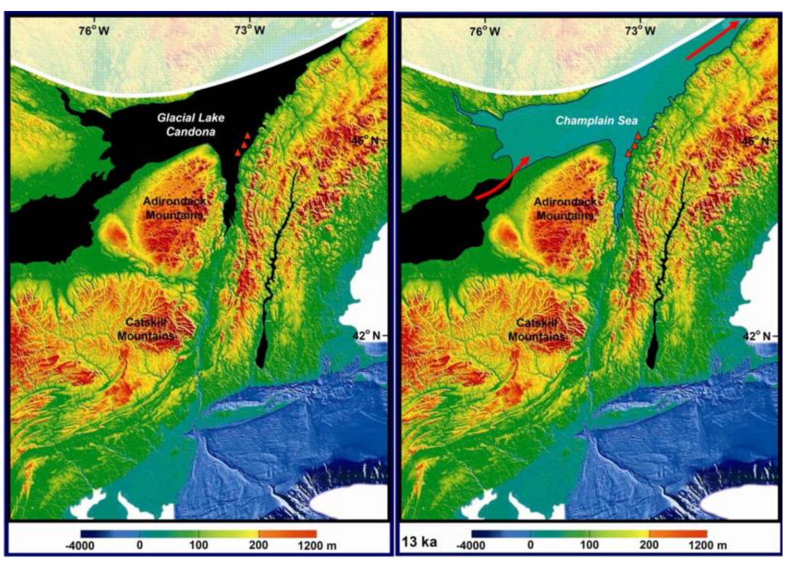
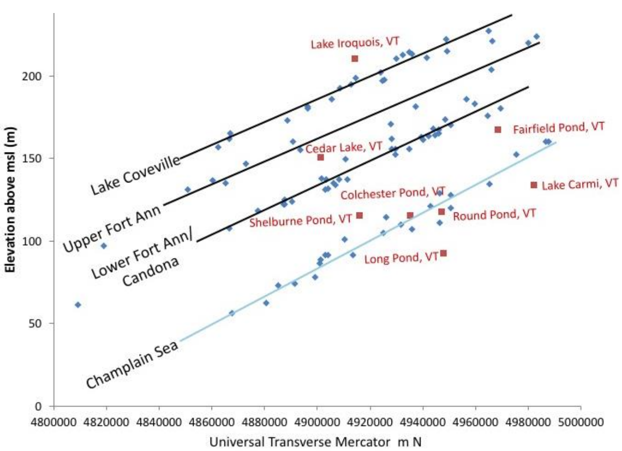
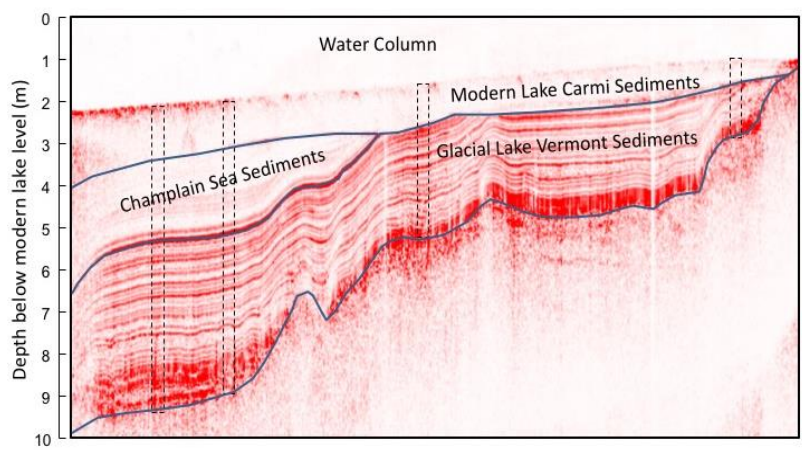


Figure 1
Maps of the extent of Glacial Lake Candona (left) and the Champlain Sea (right). The red triangles indicate coring sites for this study in Lake Carmi, Fairfield Pond, and Monkton Pond. The red arrows indicate the route of meltwater through the St. Lawrence Valley. The white area represents the Laurentide Ice Sheet; the retreat of the ice sheet opened a conduit for drainage of the lake.

We can therefore stratigraphically locate the opening of the Saint Lawrence Valley by locating the transition from glacial lake sediments to marine sediments in the Champlain lowlands. We identified three modern lakes—Lake Carmi, Fairfield Pond, and Monkton Pond—within the boundaries of Glacial Lake Candona. Because sedimentation rates are lower on lake margins than in lake interiors, we aimed to collect a long sediment core from a site that occupied the margins of the former Glacial Lake Candona basin (*Figure 1).* By recovering terrestrial macrofossils from the cores in the intervals of interest, we would be able to build an age model without the uncertainty inherent in determining reservoir corrections for marine fossils. Such terrestrial fossils are rare in ice proximal deposits, so few studies have used terrestrial material to date marine strandlines (Wagner 1972, Tremblay 2005, Rayburn et al. 2007). We aimed to find material of sufficient mass to lower the uncertainty associated with Champlain Sea inception.

Paleoshoreline reconstructions (*Figure 2)* suggest that Fairfield Pond lies at too high an elevation to have been inundated by the Champlain Sea. When the St. Lawrence Valley opened and Lake Candona drained, this site became an open basin. However, the Carmi site was covered by both the glacial lake and the inland sea. We expect a transition from glacial lake to marine sediments in the Lake Carmi cores and a transition from glacial lake to open basin sediments in the Fairfield Pond cores. We will distinguish the lacustrine and marine facies using stratigraphy and microfossils; presence of the freshwater ostracod Candona subtriangulata will be indicative of the glacial lake, and marine foraminifera of the Champlain Sea.





Figures 2 and 3

2. A significant body of work has used features such as ridges, spits, and cliffs to identify the chronology of lake phases in the Champlain lowlands; this work is reviewed by Franzi, Rayburn et al. (2002). Four phases of Glacial Lake Vermont, of which Glacial Lake Candona is the lowest, have been mapped. The reconstructed paleoshorelines indicate that the modern-day Lake Carmi basin was inundated by the Champlain Sea, but Fairfield Pond is at too high an elevation to have been inundated. *3.* Data from the CHIRP sub-bottom profiler show a dramatic transition from highly-reflective, laminated sediments to less reflective sediments. We interpreted this as the transition from Glacial Lake Candona (the final stage of Glacial Lake Vermont) to the Champlain Sea.

METHODS

In 2017, we conducted geophysical surveys of the three lakes in northern Vermont to select targets for coring. The surveys were conducted using an EdgeTech CHIRP (Compressed High Intensity Radar Pulse) sub-bottom profiler, with 10 ms sound pulses sweeping frequencies between 4 and 24 kilohertz. We interpreted the visible transition in the CHIRP data from highly-reflective, laminated sediments to less reflective sediments as the transition between the glacial lake and inland sea sediments, and identified locations where our 9-meter core barrels could penetrate the Glacial Lake Candona sequence and reach the bedrock beneath (figure 3).

In May 2018, we returned to the three lakes to recover sediment cores using our Rossfelder vibracore system. We used 3-inch aluminum barrels with installed polycarbonate core catchers for sediment recovery. We collected 7 cores of lengths ranging from x to x from Lake Carmi, 8 cores ranging from x to x in Fairfield Pond, and 3 cores ranging from x to x in Monkton Pond, for a total of x meters of sediment. Upon retrieval, the cores were sawed into sections of 150 cm of less for transport. The cores from Lake Carmi are notated CARa-Db-c:d, where a indicates the site, b indicates the drive number, and c:d indicates the section number, with section 1 at the top. The cores from Fairfield Pond are notated RON, and the cores from Monkton Pond are notated DOODY. We reached the critical stratigraphic interval in Carmi and Fairfield not in Monkton Pond, due to the centuries-long use of the pond as a repository for waste from the surrounding dairy farms. Therefore, DOODY1, 2, and 3 are excluded from further analysis.

The cores were split, imaged, and described in the Coastal Research Laboratory. Thus far, we have opened five cores from each site; two more cores remain to be opened from Lake Carmi, and three more cores from Fairfield Pond. To describe these cores, we noted visual changes in color and texture and sampled every 10 cm for texture. We identified pebbles; coarse, medium, and fine sand; sandy silt and silt; and organic-rich silt. The cores from each site followed a pattern unique to the site (*Figure 4)*.

Fairfield Pond Stratigraphy **Lake Carmi Stratigraphy**

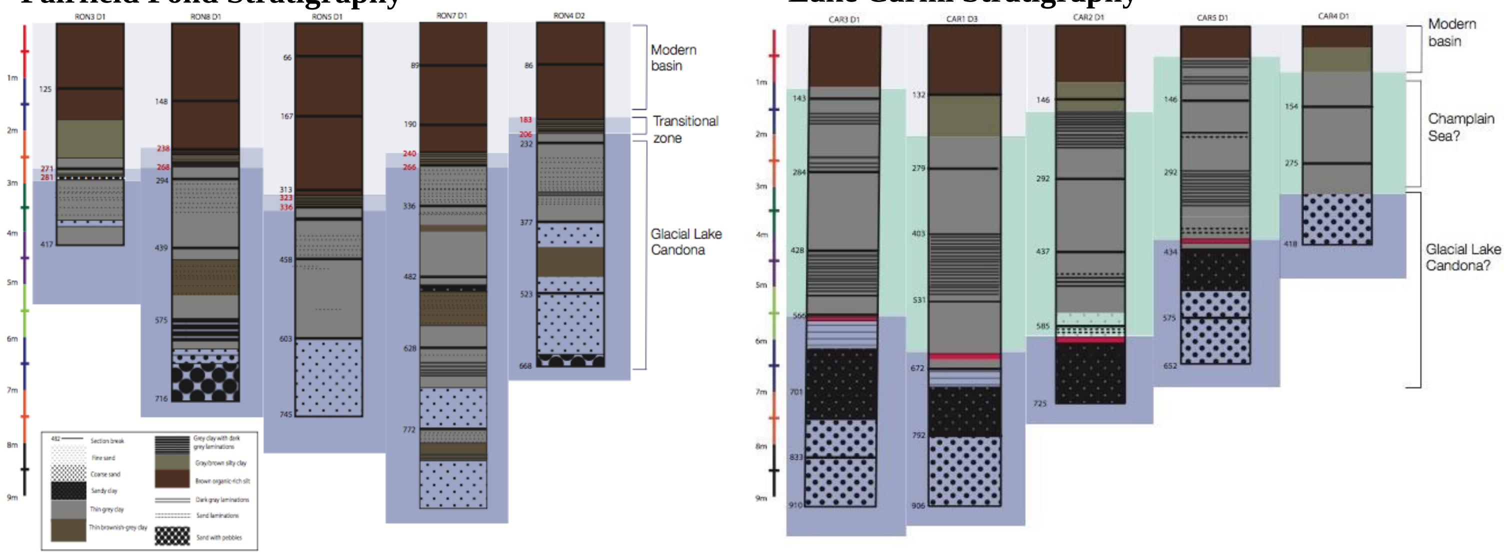


Figure 4

2. The visible transition from the grey clay, interpreted as Glacial Lake Candona sediment, to dark brown organic-rich silt, interpreted as sediment from the open basin, is dramatic. The changes in sediment throughout the laminated zone are also dramatic. This transition is confirmed by the abrupt cessation of freshwater ostracods just below the transitional zone. The stratigraphy in Lake Carmi is less legible than the stratigraphy in Fairfield Pond; the Candona-Champlain transition is not visibly apparent. Rayburn et al. (2011) identify a reddish brown clay interval at the lacustrine-marine transition, which we found in four of the five Carmi cores. We find abundant ostracods below the reddish layer and carbonate shells (at present, unidentified, but potentially foraminifera) above, so we also interpret this reddish layer as the Champlain Sea onset. We do not have an explanation for the origins of the red sediment.

In Fairfield Pond, the stratigraphy has the following trend: (1) dark brown organic-rich silt, (2) transitional zone of approximately 20 alternating 1 cm laminations (3) grey clay with medium-fine sandy laminations, (4) medium-coarse sand, (5) pebbles. In Lake Carmi, the trend is significantly different: (1) dark brown organic-rich silt, (2) grey clay, (3) grey clay with fine sand-silt laminations, (4) grey clay, (5) grey sandy clay, (6) sandy clay/coarse sand. We also noted a distinct 5 cm reddish-brown layer that appears above the sandy clay/coarse sand facies in four of the five Carmi cores. We designated the working and archive halves of the core and pulled radiocarbon samples visible on the surface of the working halves.

After splitting five cores from each site, we began to sieve the cores at 63 microns in order to recover terrestrial organic matter for radiocarbon analysis. We began with the Fairfield Pond cores, because the termination of Glacial Lake Candona was visually obvious in the transition between grey clay and brown organic-rich silt. This change occurred in each Fairfield core over an approximately 20 cm transitional zone, characterized by grey clay and dark brown silt laminations. We interpreted the first transition from clay to silt as the initial Laurentide retreat and Lake Candona drainage, and therefore targeted the intervals directly above and below this transition when sieving. We sieved in 2-centimter increments, sampling 1/3 the width of the core at a time, and picked organic matter from the captured sediments. Samples were sieved using tap water but rinsed and further treated using only deionized water.

We submitted seven samples from Fairfield Pond for radiocarbon dating (*Figure 4).* From RON4, we retrieved four samples: one from the top of the transitional zone, two from the 2 cm directly above the transition, and one from the 2 cm below. The two samples from the same depth above the transition (B1 and B2 from RON4_D2_2:5_114-116) are a twig (14.1 mg) and bulk sediments. We hoped to test if bulk sediments can be used in the absence of a single massive sample. From RON8, we retrieved three samples, two of which are bulk fragments. For sample E (RON8_D1_2:5_117-119), we separated plant and insect material before preparing for submission. All seven samples were dried overnight and photographed before being vialed. We submitted the samples to NOSAMS for radiocarbon dating. The most critical samples for constraining the date of the opening of the St. Lawrence Valley would be those from directly

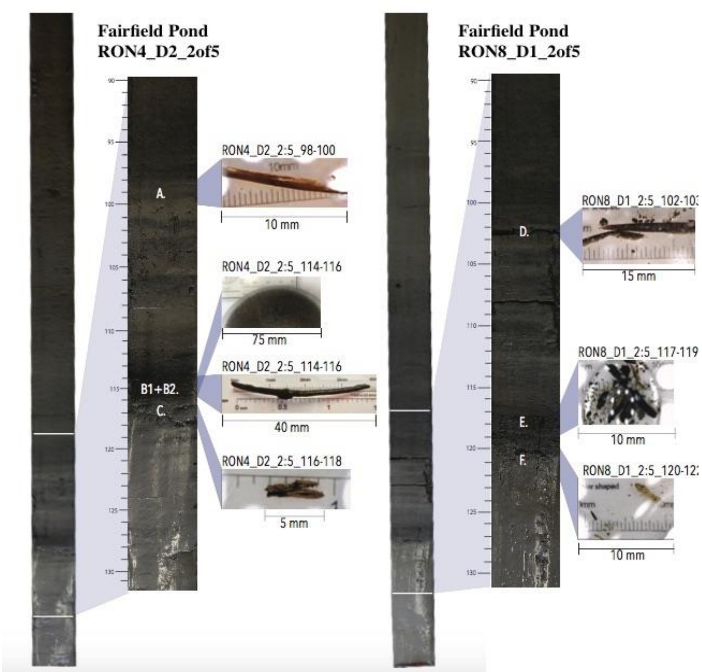


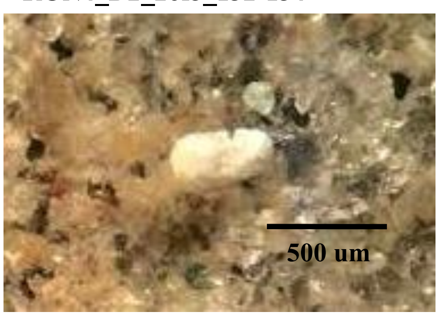
Figure 5

Sieving at 63 um throughout the laminated silt-clay transitional interval and in the clays below the transition in the Fairfield Pond cores yielded seven samples. These samples ranged in mass from 0.1 mg (sample C) to 112.3 mg (B2). Where possible, we isolated plant matter from the organic matter separated during sieving. Samples from Lake Carmi have thus far been too small to submit for dating. These samples will be critical for dating the final drainage of Lake Candona and the opening of the St. Lawrence Valley.

above and directly below the clay-silt transition: B1/B2 & C from RON4_D2, and E & F from RON8_D1. Dates for the samples from the top of the transitional interval would provide insight into the silt-clay laminations. Did this interval contain the Younger Dryas, with the clays representing periods of glacial re-advance? Or did the interval extend into the Holocene?

We then began working with the Lake Carmi cores. We anticipated using *Candona subtriangulata* to locate the Glacial Lake Candona facies, and marine foraminifera to locate the Champlain Sea facies. We began sieving 2 cm intervals in the bottom section of CAR3_D1, planning to move upwards through the sections until *Candona* terminated and foraminifera began to appear. Initially, we did not find any *Candona* or foraminifera. We interpreted the lack of foraminifera as an indication that the water was too fresh for the organisms to tolerate. However, upon finding charcoal and realizing that any pressure applied during sieving would cause the charcoal to disaggregate, we began to sieve using only water pressure (i.e. no additional hand rubbing). We then began finding ostracods in CAR3_D1_5:7 and CAR1_D3_3:5. We believe these ostracods are *Candona*, given their apparent similarity in texture and shape to the ostracods that Cronin (1977) identified as *Candona.* We also found carbonate material in CAR3_D1_4of7 that may have been foraminifera fragments, though we were not able to definitively identify them. We pulled one sample, a 1.3 mg piece of bark, from CAR3_D1_4:7, but have not yet submitted it.

RON4_D2_2of5_132-134



CAR5_D1_3of5_132-134

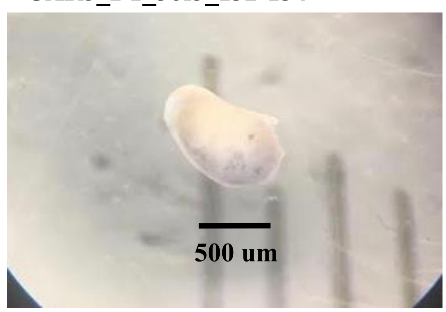


Figure 6
Ostracods found in Fairfield Pond below the silt-clay transition (left) and in Lake Carmi above the red lamination (right) appear to be different species. The ostracods in Carmi appear more similar to the ostracods Cronin identified as *Candona subtriangulata.*

RESULTS

The radiocarbon ages we received from NOSAMS were calibrated using the CALIB 7.0 program and the INTCAL13 dataset to produce a range of calendar ages for each of the seven samples (Stuiver et al., 2014). The CALIB program produces multiple possible age ranges for each sample, so we selected the age range with the highest area under the probability distribution. We considered these dates in context of the 12,896 ± 138 cal years BP Younger Dryas onset as dated in NGRIP, which is the most highly-resolved of the Greenland ice cores (Rasmussen et al, 2006). The age ranges for the samples from below and above the transition in RON4—samples B1 (11,274-11,629) and C (13,132-14,193)—bracket the Younger Dryas onset but do not constrain it. This preliminary age model confirms what we previously knew: Glacial Lake Candona existed before the Younger Dryas, and had ceased to exist by the end of the Younger Dryas.

Interestingly, the sample from immediately above the transition (B1) very closely post-dates the Younger Dryas termination, dated at 11,703 ± 99 cal years BP (Rasmussen et al., 2006). This finding may suggest that deposition ceased during the Younger Dryas, but more likely indicates that sediment deposited during the YD was subsequently eroded. These erosional

	Core	Depth (cm)	Item	Weight (mg)	Age (C14 years)	Error (years)	Age (cal years BP, 2σ)
A.	RON4 D2 2:5	98-100	Grass	0.2	8,960	130	9,655-10,401
B1.	RON4 D2 2:5	114-116	Twig	14.1	10,000	40	11,274-11,629
B2.	RON4 D2 2:5	114-116	Bulk sed	112.3	10,700	45	12,576-12,719
C.	RON4 D2 2:5	116-118	Bark	0.1	11,800	230	13,132-14,193
D.	RON8 D1 2:5	102-103	Twig	1.4	8,900	40	9,897-10,189
E.	RON8 D1 2:5	117-119	Fragments	1.9	12,850	460	13,777-16,552
F.	RON8 D1 2:5	120-122	Fragments	0.7	12,050	20	13,333-15,026
G.	CAR3 D1 4:7	102-104	Bark	0.3	-	-	-

Figure 6 (above)
Sample depth, type, radiocarbon age, and calibrated calendar age for the seven Fairfield Pond samples submitted to NOSAMS. Calendar ages are given as the range of years within two standard deviations (2 sigma). The CALIB 7.0 program produces multiple possible age ranges for each sample, so we selected the range with the highest area under the probability distribution. Samples B2, E, and F are likely driven old by the presence of reworked material.

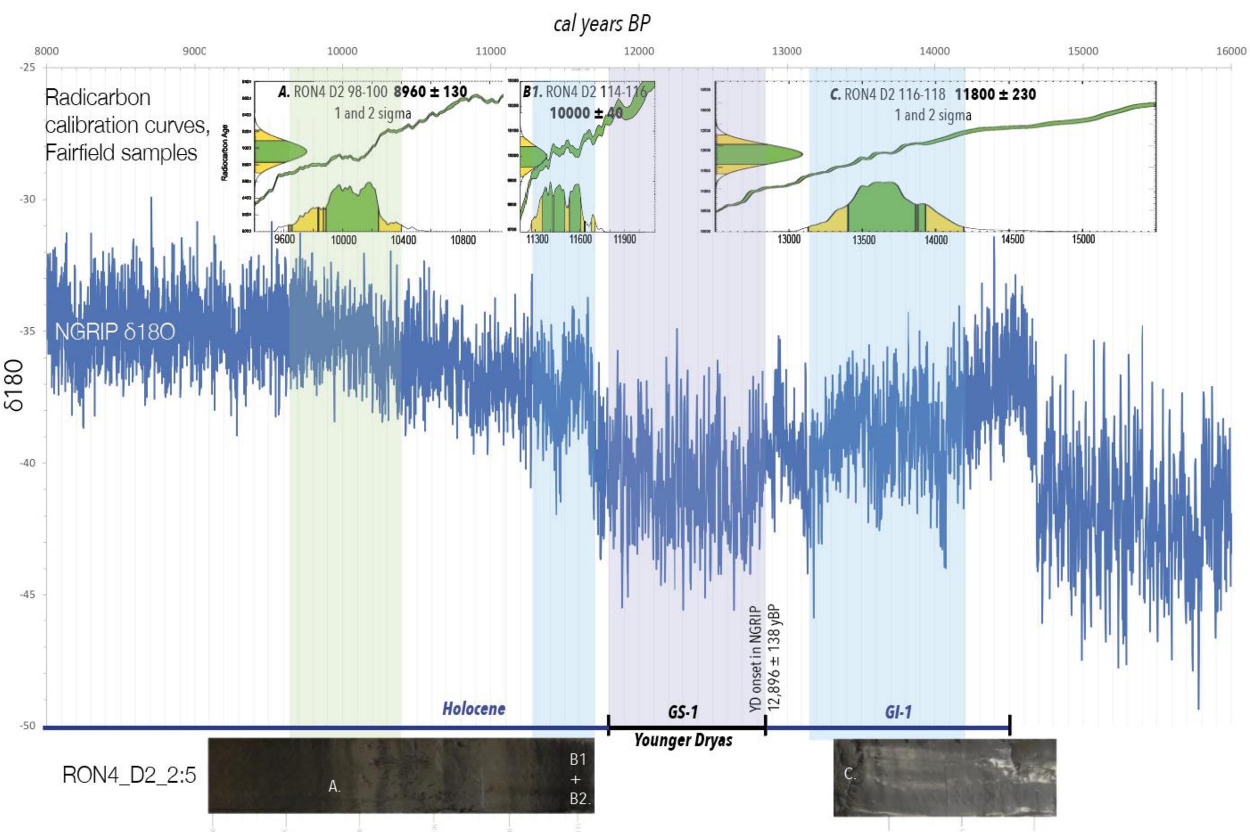


Figure 7 (above)
The radiocarbon calibration curves for the three RON (Fairfield) samples aligned with photographs of the cores NGRIP oxygen isotope record, which dates YD onset at 12,896 ± 138 cal years BP and termination at 11,703 ± 99 cal years BP. The alignment suggests that the Younger Dryas is missing from the core.

events perhaps occurred during pulses of melt following glacial re-advance during the Younger Dryas or during the abrupt warming (15˚C over decades) at the end of the cold period. We need additional dates from directly above the transition to confirm and/or constrain the duration of the missing record, in order to understand what produced the surprising unconformity.

The fact that the grass from the top of the transitional interval (Sample A from RON4_D2) was dated 9,655-10,401 cal years BP indicates that for the two millennia following YD termination, the basin was perhaps receiving fluxes of meltwater from proglacial lakes in the east. Deglaciation and uplift in the eastern Great Lakes region may have caused freshwater events that deposited glacial sediment (the gray clays) in the Fairfield basin. Or, perhaps the laminations indicate a basin alternating between two different states: a closed lake basin with high organic productivity (the brown silt) and an interconnected inland sea (gray clay). Alternatively, they could have been produced in a closed basin with alternating productivity states, with the brown silt intervals representing century-long periods of high productivity. In any case, it appears that the local climate system experienced century-scale fluctuations throughout the first two millennia of the Holocene before settling into stability around 9,000 cal years BP. Future investigation is needed into the composition and timing of the laminae to understand the end-glacial/early Holocene climate story they preserve.

We identified ostracods below the silt-clay transition in RON4_D2 that terminated abruptly at the transition. Based on Cronin's (1977) identification of Champlain Sea foraminifera and ostracods, we do not believe these ostracods to be Candona. Because these intervals were sieved using hand pressure, and any rubbing of the sediment against the sieve seemed to crush the *Candona* in the Lake Carmi cores, it is likely that the ostracods we found in Fairfield are more robust than *Candona*. The ostracods thus far identified in Fairfield Pond (*Figure 6)* have a more dimpled texture than the ostracods in Carmi. We need to re-sieve the clay below the transition in the Fairfield Pond cores to understand the chronology

We think that samples B2, E, and F do not provide reliable dates. Because these are bulk samples composed of multiple fragments (in the case of B2, of hundreds of pieces of microscopic organic material that formed a film atop the sample), they likely contain reworked material that drove the apparent age of the sample older. We therefore excluded them from the present analysis and will need to recover new samples from these intervals to build a more accurate age model.

The stratigraphy in Lake Carmi was less legible than the stratigraphy in Fairfield Pond. We were not able to identify a visible transition between glacial lake and marine sediments. However, now that we have begun to recover ostracods in Lake Carmi sediments, we have narrowed the stratigraphic window in which the transition likely takes place. Rayburn et al. (2011) report a reddish clay layer at the first lacustrine-marine transition in cores from the Champlain Valley near Plattsburgh, New York. We find a reddish layer in four of the five Carmi cores, approximately in the middle of the cores. This position aligns with the apparent transition in reflectivity in the CHIRP data. We also found abundant ostracods below and directly above the reddish layer in CAR3_D1 and CAR5_D1, decreasing in abundance over the 10 cm above the layer. Though we still need to develop an explanation for the reddish sediment, our current hypothesis is in agreement with Rayburn (2011) that this region of the core contains the transition of interest.

DISCUSSION

Rayburn et al. (2007) used terrestrial macrofossils to determine the age of the Champlain Sea incursion. They found two radiocarbon samples from the proglacial lake phase that constrain the initiation of the Champlain Sea: a musk-ox bone with a radiocarbon age of 11,362 ± 115 years (calibrated to 13,033-13,446 years BP) and a wood fragment with a radiocarbon age of 10,901 ± 76 years (calibrated to 12,691-12,973 years BP) (Stuiver et al., 2014). The wood fragment is located 1 cm above what they interpret to be sediment deposited in the drop to the Coveville level of Glacial Lake Vermont (*see Figure 2)*. They use these samples to suggest that the first floods from the draining of Glacial Lake Vermont occurred between 13,033-12,691 years. Given that the last phase of the lake, the Fort Ann Phase, lasted between 168 (Rayburn et al, 2005) and 240 years (Ridge et al, 1999), they also suggest that the initiation of the Champlain Sea may have occurred after the onset of the Younger Dryas. Rayburn therefore provides a lower bound (i.e. further back in time) to when the St. Lawrence Valley may have opened.

We hoped to provide an upper bound, or a more tightly constrained lower bound. However, we were not able to do so. Our data agree with Rayburn's data in leaving open the possibility of a later, post-YD onset opening of the St. Lawrence Valley, but do not contribute to narrowing the window of Lake Candona drainage (and thus St. Lawrence Valley opening). As stated repeatedly, we need additional radiocarbon ages from both Fairfield and Carmi to narrow this window and affix certainty to a later valley opening. Radiocarbon dates from Lake Carmi are particularly important, because we can only infer the initiation of the Champlain Sea from the Fairfield cores, and there appears to be a significant unconformity. A terrestrial fossil from the sediments at the onset of the Champlain Sea is necessary to constrain the onset of the Champlain Sea.Additional dates not only provide more chronological anchors, but help develop our understanding of the time scales of the laminations, which is necessary in order to use the laminations to quantify the duration of the glacial lake stages.

If the St. Lawrence Valley opening and initiation of the Champlain Sea were indeed after Younger Dryas onset, the debate about the YD trigger perhaps shifts back to the Mackenzie River and the Arctic Ocean (Teller 2005, Keigwin 2018). Questions also emerge about the magnitudes of the first drainages (Coveville to Upper Fort Ann, Upper Fort Ann to Candona, and Candona to Champlain Sea; *see Figure 2*) and whether these flood events contributed to AMOC reduction and cooling, as well as whether the final draining of Glacial Lake Candona—which, according to this model, occurred within the Younger Dryas—did not cause further AMOC inhibition.

FUTURE WORK

We hope that future sieving in Fairfield Pond will yield radiocarbon dates that more tightly constrain the end of Glacial Lake Candona and reveal the chronology of the laminations in the transitional zone. Now that we have begun to recover ostracods in the Lake Carmi cores, we will use carbonate microfossils to locate the Candona-Champlain transition. We hope to recover terrestrial macrofossils of sufficient mass to provide precise dates that constrain the initiation of the Champlain Sea and the opening of the St. Lawrence Valley. Additional radiocarbon samples and a more precise chronology of Glacial Lake Candona drainage (as expressed in both Carmi and Fairfield cores) and Champlain Sea onset (as expressed in Carmi) will help us understand the dynamics of lake drainage and inland sea onset. Was the exchange rapid, expressed in an abrupt transition from lacustrine to marine sediment and/or fresh to brackish species assemblages, or more gradual, expressed in a period of brackish water?

We hope that XRF scans of the cores will help us in identifying the Candona-Champlain transition in Carmi, and in interpreting the deglacial chronology at both sites. The sediments we currently understand as the Champlain Sea sequence are thinly laminated, and an XRF scan of this sequence together with a robust age model would help us count and ultimately interpret the laminations. Sediments deposited during flood events have distinct physical and chemical characteristics, so the elemental composition provided by the XRF—together with the robust age model—will help us locate the major events in the deglacial history.

To assess the role of meltwater from Glacial Lake Candona in triggering the Younger Dryas, we will use a digital elevation model of the Champlain lowlands to estimate the volume of water stored in Glacial Lake Candona. Efforts have been made to quantify the volume of water released during the Coveville-Upper Fort Ann discharge (Franzi et al., 2002), but not the Upper-Lower Fort Ann/Candona or the Fort Ann/Candona-Champlain Sea discharges (*see Figure 2).* We will use strandlines visible in the DEM and known rates of isostatic rebound to quantify the volume of water that would have drained through the St. Lawrence upon retreat of the LIS, and input these volumes into models that predict impacts of freshwater fluxes on AMOC.

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