Synchronizing a sea-level jump, final Lake Agassiz drainage, and abrupt cooling 8200 years ago

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Abstract

Freshwater pulses draining into the North Atlantic Ocean are commonly hypothesized to have perturbed the Atlantic meridional overturning circulation (MOC), triggering abrupt climate changes such as Heinrich events, the Younger Dryas, and the 8.2 ka event. However, dating uncertainties have prevented causal links between freshwater pulses and climate events from being firmly established. Here we report a high-resolution relative sea-level record from the Mississippi Delta that documents a sea-level jump that occurred within the 8.18 to 8.31 ka (2σ) time window and is attributed to the final drainage of proglacial Lake Agassiz–Ojibway (LAO). This age is indistinguishable from the onset of the 8.2 ka climate event, consistent with a nearly immediate ocean–atmosphere response to the freshwater perturbation. This constitutes a rare currently available example of a major abrupt climate cooling that can be directly linked to a well-documented freshwater source with a temporal resolution on the order of a century. The total inferred eustatic sea-level rise associated with the very final stage of LAO drainage at 8.2 ka ranges from 0.8 to 2.2 m, considerably higher than previous estimates. These new constraints on the timing and amount of final LAO drainage permit significantly improved quantitative analysis of the sensitivity of MOC to freshwater perturbation, a crucial step toward understanding abrupt climate change.

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1. Introduction

Abrupt climate change has received extensive interest for a wide range of reasons, including its potential role in a future warming world (Alley et al., 2003). Over the past few decades, the connection between freshwater forcing and abrupt climate change due to perturbation of the Atlantic meridional overturning circulation (MOC) has enjoyed widespread popularity, since it offers a potential mechanism to explain phenomena such as Heinrich events (Heinrich, 1988), the Younger Dryas (Broecker et al., 1989), and the 8.2 ka event (Barber et al., 1999). However, the past few years have seen this hypothesis becoming increasingly challenged (e.g., Barber et al., 2010; Fisher et al., 2008; Lowell et al., 2009), in part reflecting the fact that very few abrupt climate events have been unequivocally linked to a well-dated freshwater source (cf. Clement and Peterson, 2008).

The 8.2 ka cold event is the most prominent abrupt North Atlantic climate change of the Holocene and is increasingly recognized in many other parts of the world (Alley and Ágústsdóttir, 2005; Cheng et al., 2009). This event is often believed to have resulted from the final outburst of proglacial Lake Agassiz–Ojibway (LAO) when an ice dam over Hudson Bay collapsed (Barber et al., 1999; Lajeunesse and St-Onge, 2008) and the rapid drainage flooded the North Atlantic Ocean with freshwater and perturbed the Atlantic MOC (Ellison et al., 2006; Kleiven et al., 2008), leading to widespread cooling. In addition, the rerouting of western Canadian Plains runoff following the collapse of the ice dam over Hudson Bay may have contributed to the 8.2 ka climate event (Carlson et al., 2009). Despite the popularity of a causal link between the final LAO drainage and the 8.2 ka climate event, this relationship has yet to be firmly demonstrated because the catastrophic LAO drainage remains poorly constrained in terms of its timing and amount. The only available direct dating of the final LAO drainage yields an age range of 8.16 to 8.74 ka at the 1σ level (Barber et al., 1999). This large age uncertainty precludes an unequivocal relationship has yet to be firmly demonstrated because the catastrophic LAO drainage remains poorly constrained in terms of its timing and amount. The only available direct dating of the final LAO drainage yields an age range of 8.16 to 8.74 ka at the 1σ level (Barber et al., 1999). This large age uncertainty precludes an unequivocal connection between LAO drainage and the 8.2 ka event and allows for alternative hypotheses such as a role for solar forcing around this time interval (Muscheler et al., 2004; Rohling and Pälike, 2005). Also, the amount of LAO drainage is not well known as reflected by highly variable estimates (e.g., Barber et al., 1999; Hjima and Cohen, 2010; Leverington et al., 2002; Törnqvist et al., 2004a), inhibiting our understanding of the sensitivity of MOC to freshwater perturbation.

This study seeks to refine previous work (Törnqvist et al., 2004a) that provided the first evidence for a sea-level jump around 8.2 ka based on stratigraphic data from the Mississippi Delta, Louisiana, USA.
We present a high-resolution relative sea-level (RSL) record around this time interval using basal peat to track sea-level change. The rationale of this approach is that rising seas drown the coastal landscape and transform it into a peat-forming wetland that accumulates over a consolidated, compaction-free Pleistocene basement. Therefore, intertidal basal peats can be used to determine past sea levels with high accuracy via precise measurements of their age and elevation. The robustness of this approach has been demonstrated in a variety of coastal settings (e.g., Donnelly et al., 2004; Jelgersma, 1961).

2. Study area

Coastal plains worldwide (e.g., the US Atlantic Coast) rarely capture the age/depth range necessary to sample early Holocene sea-level records that are more likely found in large, prograding deltas. However, not all deltas contain basal peat and even fewer also occur in microtidal settings which are particularly favorable for high-resolution sea-level studies. Our sampling sites are located in the Bayou Sale area in the western part of the Mississippi Delta (Fig. 1). The US Gulf Coast is characterized by a microtidal regime with a present-day spring tidal range typically <0.5 m in coastal Louisiana. In addition, the study area has been tectonically relatively stable during the Holocene (Törnqvist et al., 2006). Glacial isostatic adjustment (GIA) contributes significantly to RSL rise in this area around 8.2 ka (Kendall et al., 2008), but the GIA component would be negligible during a short-lived sea-level jump. This overall combination of circumstances makes our study area exceptionally well suited to resolve dm-scale RSL changes for this time interval.

The Pleistocene basement in the study area consists of the pervasively oxidized Prairie Complex (Autin et al., 1991) that is capped by a few meters of Peoria Loess. Both units are highly consolidated and essentially compaction-free due to prolonged subaerial exposure. Overlying the Peoria Loess is an immature paleosol consisting of an A-horizon enriched in highly decomposed organic matter. This paleosol was classified as an Entisol, suborder Aquent, by Törnqvist et al. (2004b) and is the result of transgression, a rising groundwater table, and the initial transformation of the landscape into a wetland environment. The continued rise of the groundwater table eventually enabled the formation of basal peat. The distinction between the paleosol and the basal peat is based on (1) the dark gray matrix color for the paleosol vs. gray brown for the basal peat; (2) the lesser degree of organic matter decomposition in the peat as reflected by abundant herbaceous plant fibers; and (3) the massive structure of the paleosol compared to the faintly laminated peat. Nevertheless, basal peat can have a significant mud content and occasionally contains distinct mud beds.

3. Methods

We collected cores with a Geoprobe system (model 6610 DT). The early stage of coring aimed at mapping the stratigraphy along a ~6-km-long transect (Fig. 1), exhibiting a transgressive surface associated with the Pleistocene–Holocene transition. Subsequent efforts were focused on coring at key locations for detailed sampling to improve the precision of depth measurements of this transgressive surface.

Cores were initially described in the field and then transported to Tulane University for cold storage (~4 °C). In the laboratory, representative cores containing basal peat were sampled for radiocarbon dating, carbon isotope measurements, and foraminiferal analysis to determine the chronology of basal peat and to constrain depositional environments of both basal peat and adjacent strata. Radiocarbon dating of terrestrial plant remains from basal peat was performed by accelerator mass spectrometry (AMS) at the University of California, Irvine. Stable carbon isotope and foraminiferal analyses of two representative cores (sites Bayou Sale VI and IV) were performed to characterize depositional environments. For δ13C analysis, samples were first dried at 60 °C for 24 h and acidified with 10% HCl to remove carbonates. The residues were centrifuged and the isolated organic material was then dried overnight at 60 °C. δ13C measurements were carried out at the Stable Isotope Laboratory at the University of Miami. For the foraminiferal analysis, samples were soaked in water for 24 h, wet sieved, and the >63 μm fraction was examined under a microscope. Identification of agglutinated foraminifera was based mainly on pseudo-chitinous linings because complete outer tests were often lacking due to poor preservation.

Optical surveys with an infrared TOPCON GTS-4B total station were conducted between core sites and National Geodetic Survey (NGS) benchmark T168 (UTM-coordinates: N = 3281.840; E = 645.980) (Fig. 1) to determine the land surface elevation at the core sites. In addition, temporary benchmarks were established between the NGS benchmark and core sites. The temporary benchmarks (not shown in Fig. 1) were located very close (typically = 100 m) to the core sites. At least two round-trip surveys were carried out between the NGS benchmark and temporary benchmarks, and typically two round-trip surveys were conducted between a temporary benchmark and a core site. The cumulative error for a round-trip elevation survey between the NGS benchmark and core sites is within 0.05 m.

4. Results

4.1. Stratigraphy

We drilled 37 sites along the ~6-km-long transect to map the stratigraphy in the Bayou Sale area (Fig. 1); key stratigraphic information for all core sites is summarized in Table 1. Multiple cores that capture the Pleistocene–Holocene transition were drilled at the majority of the sites.

The transgressive succession at the stratigraphically deeper sites (V, 32, VII, and VI) is characterized by a basal-peat bed that caps the dark gray paleosol described above and is abruptly overlain by pale-gray, shell-bearing muds (Figs. 2, 3). The basal-peat bed at the deepest sites (V, 32 and VII) shows highly variable characteristics and thicknesses among multiple cores at each site and is often absent in this deeper portion of the record due to erosion (Table 1). This is...

3.1. Depositional environment

The depositional environments associated with the facies described above are reconstructed by means of foraminiferal and stable carbon isotope analysis. Fig. 4a shows the succession of foraminiferal assemblages and other microfossils at site Bayou Sale VI. The basal peat bed and the underlying paleosol are dominated by the agglutinated taxa *Haplophragmoides wilberti* and *Ammoastuta inepta*. This interval is interpreted to represent a brackish marsh environment. In the mud above the basal peat, the microfauna is dominated by calcareous foraminifera of the taxa *Ammonia beccarii* sl. and *Elphidium gunteri*, with *Ammobaculites* spp. being the next dominant genus along with occurrences of *H. wilberti*. *Ammobaculites* spp. is represented only by the early coiled portion of the test and therefore no species identification was possible. The calcareous foraminifera (*Ammonia* and *Elphidium*) occur in open water where salinities are generally greater than 10 ppt (Kane, 1967). The interval above the peat is therefore interpreted to represent an open-water, brackish lagoon environment. Fragments of pelecypod taxa *Rangia cuneata* and *Macoma mitchelli*, characteristic of shallow brackish environments with salinities of 2–15 ppt (Parker, 1959; Phleger, 1965; LaSalle and de la Cruz, 1985) occur just above the basal peat (Fig. 1; sites VI, VII, and 32), providing additional evidence that the brackish marsh was abruptly replaced by a brackish lagoon.

Fig. 4b shows the succession of foraminiferal taxa and other microfossils at site Bayou Sale IV. The section below 14.0 m is dominated exclusively by *H. wilberti* and interpreted as a brackish marsh environment. The interval above 14.0 m is represented by a *H. wilberti–* *A. inepta* assemblage and is also interpreted as a brackish marsh, possibly with a lower salinity due to the occurrence of *A. inepta*. Scott et al. (1991) recorded *A. inepta* in Louisiana marshes with salinities ranging from 3 to 5 ppt. *H. wilberti* and *A. inepta* are represented in most samples only by their pseudochinous linings. In one sample (13.76 m) whole specimens with the fragile test intact were preserved, allowing for positive identification of the species and associated linings. The general environmental setting is similar to that described by Kane (1967) where a *H. wilberti–* *A. inepta* assemblage occurs as part of a fringe marsh with salinities less than 10 ppt.

The basal peat at sites VI and IV yielded δ13C values of −13.0% and −12.9 to −15.6%, respectively, also indicative of a brackish marsh environment (Chmura et al., 1987). The combined micropaleontological and geochemical data provide conclusive evidence that the basal peat at both sites accumulated within the intertidal zone (between mean tide level and mean spring high water). While at site IV this environment persisted up section, at site VI the marsh was abruptly replaced by a brackish lagoon. Given the straightforward relationship between microfossil content and lithofacies, all cores presented in this study (Table 1, Fig. 1) can be readily interpreted in terms of depositional environments.

4.3. Elevation and sea-level relationship of basal peat

The elevation of past sea level was calculated using depth measurements, elevation surveys, and the vertical indicative range (sensu Van de Plassche, 1986) of basal peat with respect to sea level. The depth is defined as the contact between the basal-peat bed and the underlying paleosol. Since basal peats were deposited on the highly consolidated Pleistocene substrate, this essentially eliminates elevation errors induced by post-depositional compaction. Multiple cores were collected from each site to determine the measurement error of the depth level of basal-peat beds (Table 2).
The indicative range of basal peat refers to the vertical interval in which basal-peat formation takes place with respect to mean sea level. Van de Plassche (1982) showed that basal-peat accumulation in coastal settings often occurs between mean sea level and mean high water. As shown by the brackish signature of the $\delta^{13}$Ca and $\delta^{15}$N foraminiferal data discussed above, our basal-peat samples formed...
within the intertidal zone. The average present-day spring tidal range in coastal Louisiana is 0.47 m (González and Törnqvist, 2009). Assuming the early Holocene tidal range was comparable to the modern tidal range, the indicative range of the basal peats in our study area would be 0.24 m (cf. González and Törnqvist, 2009). We convert this value to a “two-sided error” of ±0.12 m to be consistent with error designations for depth and elevation measurements.

The cumulative uncertainty of the sea-level elevation inferred from basal peat at each site can be computed with the following equation:

\[
E = \sqrt{E_d^2 + E_i^2 + E_e^2}
\]
Table 2
Summary of the elevation measurements and uncertainties of the basal peat/paleosol contact at 14C dated sites.

<table>
<thead>
<tr>
<th>Site</th>
<th>Mean elevation (m)</th>
<th>Number of measurements</th>
<th>$E_D$ (±m)</th>
<th>$E_e$ (±m)</th>
<th>$E_r$ (±m)</th>
<th>$E$ (±m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bayou Sale IV</td>
<td>−14.08</td>
<td>10</td>
<td>0.09</td>
<td>0.05</td>
<td>0.12</td>
<td>0.16</td>
</tr>
<tr>
<td>Bayou Sale V</td>
<td>−16.07</td>
<td>10</td>
<td>0.10</td>
<td>0.05</td>
<td>0.12</td>
<td>0.16</td>
</tr>
<tr>
<td>Bayou Sale VI</td>
<td>−14.41</td>
<td>7</td>
<td>0.10</td>
<td>0.05</td>
<td>0.12</td>
<td>0.16</td>
</tr>
<tr>
<td>Bayou Sale VII</td>
<td>−15.53</td>
<td>3</td>
<td>0.05</td>
<td>0.05</td>
<td>0.12</td>
<td>0.14</td>
</tr>
</tbody>
</table>

where $E$ is the total uncertainty; $E_D$ is the depth measurement error; $E_e$ is the surveying error; and $E_r$ is the indicative range error. The elevation measurements and uncertainties are summarized in Table 2.

Table 3
Radiocarbon ages of basal peat from the present study in the Bayou Sale area.

<table>
<thead>
<tr>
<th>Sample name</th>
<th>UTM coordinates¹</th>
<th>Surface elevation (m)</th>
<th>Depth below surface (m)</th>
<th>Material dated</th>
<th>UCIAMS² Lab number</th>
<th>Radiocarbon age</th>
<th>Calibrated age (cal yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(N) (E)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(14C yr BP ± 1σ)</td>
<td>Weighted mean (± 1σ)</td>
</tr>
<tr>
<td></td>
<td>(m)</td>
<td></td>
<td>(±m)</td>
<td></td>
<td></td>
<td>Phase</td>
<td>Weighted mean</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>IV-1</td>
<td>8210</td>
</tr>
<tr>
<td>IV-1</td>
<td>3279.02</td>
<td>643.58</td>
<td>0.31</td>
<td>11 Scirpus spp.</td>
<td>51101</td>
<td>7435 ± 15</td>
<td>7440 ± 10</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>25 herbaceous</td>
<td>51102</td>
<td>7840 ± 15</td>
<td>7840 ± 15</td>
</tr>
<tr>
<td>IV-1b</td>
<td></td>
<td></td>
<td></td>
<td>15 Scirpus spp.</td>
<td>51103</td>
<td>7560 ± 15</td>
<td>7580 ± 15</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>IV-2</td>
<td>8205</td>
</tr>
<tr>
<td>IV-2a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>IV-3</td>
<td>8180</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>9 herbaceous</td>
<td>51104</td>
<td>7540 ± 15</td>
<td>7540 ± 15</td>
</tr>
<tr>
<td>IV-2b</td>
<td></td>
<td></td>
<td></td>
<td>15 Scirpus spp.</td>
<td>51105</td>
<td>7315 ± 15</td>
<td>7325 ± 15</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>7 Scirpus spp.</td>
<td>51106</td>
<td>7360 ± 10</td>
<td>7360 ± 10</td>
</tr>
<tr>
<td>IV-3a</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>IV-1</td>
<td>8210</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>IV-2</td>
<td>8205</td>
</tr>
<tr>
<td>IV-3b</td>
<td>3279.40</td>
<td>644.76</td>
<td>0.33</td>
<td>9 Scirpus spp.</td>
<td>51107</td>
<td>7270 ± 60°</td>
<td>7745 ± 15</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>14 herbaceous</td>
<td>51108</td>
<td>7745 ± 15</td>
<td>7745 ± 15</td>
</tr>
<tr>
<td>V-1</td>
<td>3279.40</td>
<td>644.76</td>
<td>0.33</td>
<td>17 Scirpus spp.</td>
<td>51109</td>
<td>7525 ± 15</td>
<td>7590 ± 10</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>9 herbaceous</td>
<td>51110</td>
<td>7650 ± 15</td>
<td>7650 ± 15</td>
</tr>
<tr>
<td>V-2</td>
<td>3279.40</td>
<td>644.76</td>
<td>0.33</td>
<td>15 Scirpus spp.</td>
<td>51111</td>
<td>7690 ± 15</td>
<td>7690 ± 15</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>7 Scirpus spp.</td>
<td>51112</td>
<td>7670 ± 25</td>
<td>7670 ± 25</td>
</tr>
<tr>
<td>V-3</td>
<td>3281.10</td>
<td>646.08</td>
<td>0.55</td>
<td>24 Scirpus spp.</td>
<td>51113</td>
<td>7300 ± 15</td>
<td>7395 ± 10</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>4 large herbaceous</td>
<td>51114</td>
<td>7430 ± 15</td>
<td>7430 ± 15</td>
</tr>
<tr>
<td>V-3a</td>
<td>3281.10</td>
<td>646.08</td>
<td>0.55</td>
<td>30 small herbaceous</td>
<td>51115</td>
<td>7450 ± 15</td>
<td>7450 ± 15</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>30 small herbaceous</td>
<td>51115</td>
<td>7450 ± 15</td>
<td>7450 ± 15</td>
</tr>
<tr>
<td>V-7</td>
<td>3279.92</td>
<td>644.98</td>
<td>0.41</td>
<td>1 large unidentified</td>
<td>596574</td>
<td>7710 ± 35</td>
<td>7710 ± 25</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>2 Scirpus spp.</td>
<td>59675</td>
<td>7705 ± 35</td>
<td>7705 ± 35</td>
</tr>
<tr>
<td>VII-1</td>
<td>3279.92</td>
<td>644.98</td>
<td>0.41</td>
<td>9 charcoal fragments</td>
<td>59675</td>
<td>7705 ± 35</td>
<td>7705 ± 35</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>15 Scirpus spp.</td>
<td>59676</td>
<td>7665 ± 35</td>
<td>7665 ± 35</td>
</tr>
<tr>
<td>VII-1c</td>
<td>3279.92</td>
<td>644.98</td>
<td>0.41</td>
<td>9 charcoal fragments</td>
<td>59677</td>
<td>7650 ± 120</td>
<td>7650 ± 120</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>11 Scirpus spp.</td>
<td>59678</td>
<td>7600 ± 25</td>
<td>7605 ± 20</td>
</tr>
<tr>
<td>VII-2c</td>
<td>3279.92</td>
<td>644.98</td>
<td>0.41</td>
<td>9 charcoal fragments</td>
<td>59679</td>
<td>7610 ± 25</td>
<td>7610 ± 25</td>
</tr>
<tr>
<td>Bayou Sale</td>
<td></td>
<td></td>
<td></td>
<td>30 small herbaceous</td>
<td>7645 ± 15</td>
<td>8420</td>
<td>8400–8450</td>
</tr>
<tr>
<td>VII-2c</td>
<td>3279.92</td>
<td>644.98</td>
<td>0.41</td>
<td>30 small herbaceous</td>
<td>7645 ± 15</td>
<td>8420</td>
<td>8400–8450</td>
</tr>
</tbody>
</table>

4.4. Chronology

For each basal-peat bed, different types of terrestrial botanical macrofossils from mostly 2-cm-thick peat intervals were selected for 14C dating. We obtained 21 AMS 14C ages from sites Bayou Sale IV, V, VI, and VII (Table 3). Since cores from sites V and VII show a highly variable stratigraphy within a short distance and some cores (e.g., core D in Fig. 3) even display erosional features, utmost caution was exercised and only well-preserved basal peats were chosen for 14C dating. One 14C measurement (Bayou Sale V-1a) was rejected because it provided a younger age than all stratigraphically higher samples.

The remaining 14C ages were calibrated to calendar years Before Present (BP = AD 1950) using OxCal (v4.0) (Bronk Ramsey, 1995) and

¹ UTM coordinates (UTM zone 15R) with reference to North American Datum of 1983 (NAD83).
² UCIAMS = University of California, Irvine, accelerator mass spectrometry; Weighted means were obtained with the “combination” function of OxCal (v4.0) (Bronk Ramsey, 1995).
³ Calibrated ages shown in italic were obtained with OxCal by treating each basal peat bed independently without considering their stratigraphic order. For the OxCal sequence analysis approach, the stratigraphic order of basal peat beds is taken into account and a typical 2 cm interval within a basal-peat bed is considered a ‘phase’ for calibration (VII-1, 1 cm thick).
⁴ V-1a is rejected and calibration for V-1 was thus based on V-1b only; Calibrated ages are rounded to the nearest 5 years.
the IntCal09 calibration curve (Reimer et al., 2009). Since each basal-peat bed contains multiple $^{14}$C ages, we derived calibrated ages using the combination feature of OxCal that calculates weighted mean $^{14}$C ages prior to calibration (Table 3). Together with sites I and II from previous studies in the Bayou Sale area (Törnqvist et al., 2004a, 2006), these calibrated ages were used to reconstruct the RSL history for a ~600 yr time span around 8.2 ka (Fig. 6a). In addition, we used the OxCal sequence analysis feature that takes into account the stratigraphic order of basal-peat beds by means of a model scheme (Fig. 5). A quantitative measure of how well the calibrated ages agree with the model scheme is indicated by the “A index”. Calibrated ages with an A index over 60% are considered reliable (Bronk Ramsey, 1995). The calibrated ages of peat beds at sites IV through VII are shown in Table 3; the calibrated ages of peat beds at sites I and II are grouped together to be considered as one sequence.

### Table 4
Radiocarbon ages of basal peat from previous studies (Törnqvist et al., 2004b, 2006) in the Bayou Sale area.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Surface elevation (m)</th>
<th>Depth below surface (m)</th>
<th>Vertical Error (m)</th>
<th>Radiocarbon age ($^{14}$C yr BP ± 1σ)</th>
<th>Weighted mean ± 1σ</th>
<th>Phase</th>
<th>Calibrated age (cal yr BP)</th>
<th>A index (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bayou Sale I-1</td>
<td>0.27</td>
<td>11.56–11.58</td>
<td>0.33</td>
<td>6997 ± 40</td>
<td>6995 ± 40</td>
<td>I-1</td>
<td>7865</td>
<td>7755–7940</td>
</tr>
<tr>
<td>Bayou Sale II</td>
<td>0.48</td>
<td>13.53–13.55</td>
<td>0.35</td>
<td>7480 ± 110</td>
<td>7280 ± 30</td>
<td>II-1</td>
<td>8075</td>
<td>8010–8145</td>
</tr>
<tr>
<td>Bayou Sale II-1a</td>
<td></td>
<td></td>
<td></td>
<td>7265 ± 30</td>
<td></td>
<td>II-1</td>
<td>8075</td>
<td>8010–8145</td>
</tr>
<tr>
<td>Bayou Sale II-2a</td>
<td></td>
<td></td>
<td></td>
<td>7315 ± 60</td>
<td>7290 ± 25</td>
<td>II-2</td>
<td>8100</td>
<td>8025–8170</td>
</tr>
<tr>
<td>Bayou Sale II-1b</td>
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</tbody>
</table>

Calibrated ages from sites I and II of the previous studies were obtained in the same way as those from sites in this study.

5. Discussion

5.1. Identifying a sea-level jump

While the entire data set (Table 1) exhibits evidence of transgression and RSL rise, only one portion of the record (including sites 25 and 29; Figs. 2, 3) features open-water lagoonal muds that conformably onlap the paleosol with no basal peat. Collectively, sites IV, 25, 29, and IV record an abrupt flooding event that is unlike anything seen elsewhere in our record (Table 1). The sharp transition from basalt peat to lagoonal mud at the deeper elevation of site VI marks the onset of this flooding event, while the re-emergence of basal peat at the shallower elevation of site IV registers its end. The absence of basal peat between these two elevations at sites 25 and 29 represents the flooding event itself, when rapidly rising seas prevented coastal marsh from developing and caused direct (conformable) deposition of lagoonal mud over the underlying paleosol. The stratigraphy at sites 25 and 29 is distinctly different from the remainder of the record (Table 1) and suggests a short pulse of near-instantaneous flooding due to extremely rapid sea-level rise. It is unlikely that the distinctive stratigraphy of sites VI, 25, 29, and IV resulted from gradual RSL rise or normal faulting. Had sea level risen gradually, basal peat would occur at sites 25 and 29 as well. Furthermore, recent work (Törnqvist et al., 2006) has shown that fault activity within the study area during the Holocene has been minor.

The OxCal combination approach shows that the basal-peat beds at sites VI and IV yield indistinguishable ages of 8175–8305 and 8180–8310 (2σ) cal yr BP (Fig. 2, Table 3), respectively, indicating that this flooding event occurred within the 8.18–8.31 ka time window. The OxCal sequence analysis approach provides almost similar ages for the basal-peat bed at site VI and the lowermost 2-cm interval (IV-1) of the basal-peat bed at site IV of 8185–8310 and 8180–8255 (2σ) cal yr BP, respectively (Fig. 2, Table 3). This similarity shows that our timing of 8.18–8.31 ka for the sea-level jump is robust.

The mean elevation difference of the basal-peat beds at sites VI and IV is 0.33 m (Table 2) and the associated uncertainty was calculated following

$$\Delta E = \sqrt{E_{VI}^2 + E_{IV}^2}$$

where $E_{VI}$ and $E_{IV}$ are the total uncertainty of the inferred sea level at sites VI and IV, respectively. Since $E_{VI} = E_{IV} = 0.16$ m, $\Delta E = 0.23$ m. Therefore, the magnitude of the sea-level jump recorded between sites VI and IV is 0.33 ± 0.23 m.

5.2. Final Lake Agassiz–Ojibway drainage

The reconstructed early Holocene RSL history (Fig. 6a) suggests slightly higher rates of RSL rise before than after the sea-level jump.
of near-instantaneous sea-level rise may have punctuated this phase of the Mississippi Delta. The sharp transition from basal peat to lagoonal mud at sites V, VII, and VI resulted from sudden LAO drainage. The present study reduces the age of the final LAO drainage (Fig. 6b). It is important to note that this 130 year time interval arises from the radiocarbon calibration procedure and is thus merely associated with the intrinsic limitations of the dating technique. In addition to the stratigraphic evidence for near-instantaneous drowning, hydraulic modeling has suggested that the flooding associated with the final LAO outburst would have lasted for as little as six months (Clarke et al., 2004). Therefore, this sea-level jump must have occurred as a brief event at any time between 8.18 and 8.31 ka, not as a gradual flooding that persisted for up to 130 years.

The LAO drainage is often believed to have taken place in at least two steps (Dominguez-Villar et al., 2009; Ellison et al., 2006; Leverington et al., 2002). A high-resolution marine record from the North Atlantic reveals two distinct episodes of surface ocean freshening and associated cooling at 8.18–8.34 ka and ~8.49 ka, respectively, suggesting two pulses of freshwater discharge (Ellison et al., 2006). The striking concordance in the timing of the 8.18–8.31 ka sea-level jump and the 8.18–8.34 ka climate anomaly in the North Atlantic suggests that the sea-level jump very likely corresponds to the younger pulse of the LAO drainage (i.e., the final stage of LAO drainage). Since RSL rise prior to the sea-level jump in our study area occurred too rapidly for brackish marsh to be sustained, we cannot rule out the presence of earlier pulses of LAO drainage. For example, the similar ages of basal-peat beds at sites V and VII (Fig. 6a) may indicate such an earlier pulse of freshwater drainage around 8.4 ka, which could potentially correspond to the earlier pulse of ~8.49 ka reported by Ellison et al. (2006). (It should be noted that their age estimate is based on interpolation of limited radiocarbon dating evidence.) We also note that our oldest two samples are consistent with the age of the onset of a sea-level jump (8.54–8.38 ka, 2σ) recently recognized in the Rhine–Meuse Delta (Hijma and Cohen, 2010). Such an earlier freshwater pulse may have pre-conditioned the ocean–atmosphere system (Wiersma and Jongma, 2009), setting the stage for the principal climate event triggered by the final stage of LAO drainage. Our interpretation of a sea-level jump resulting from the final LAO drainage is also consistent with a reconstruction of the properties of the Atlantic inflow that exhibits a pronounced, abrupt freshening of the sub-thermocline at 8.2 ka, interpreted to result from glacial freshwater discharge (Thornalley et al., 2009).

5.3. Volume of the freshwater drainage

The elevation data for sites VI and IV show that the sea-level jump amounted to 0.33 ± 0.23 m (Fig. 2). A maximum of 1.2 ± 0.2 m of abrupt sea-level rise was previously estimated in the study area (Törnqvist et al., 2004a) and was subsequently considered to be dominated by glacial isostatic adjustment (GIA) (Kendall et al., 2008) which is now confirmed by our refined RSL record. These previous studies lacked the stratigraphic details that are currently available (particularly the abrupt flooding evidence from sites 25 and 29) and while GIA was indeed a significant contributor to the overall high rates of early Holocene RSL rise in this region, it was not an appreciable factor for the short-lived sea-level jump identified here.

The sea-level rise of 0.33 ± 0.23 m would mathematically define a range of 0.10 to 0.56 m for the sea-level jump, a value that must be viewed in conjunction with ecological information on marsh resiliency. It is unlikely that 0.1 m of sudden sea-level rise would leave such a widespread stratigraphic signature. Studies of modern
coastal ecosystems in the Mississippi Delta (Sasser, 1977) have shown that Scirpus spp.-dominated marshes (i.e., comparable to our reconstructed brackish marsh paleoenvironment) occur in the upper portion of the tidal frame and are flooded much less frequently compared to Spartina alterniflora-dominated salt marshes (~20 to 160 vs. ~300 times per year, respectively). In other words, given the average present-day spring tidal range for coastal Louisiana of 0.47 m, a substantial sea-level rise is needed to permanently convert a brackish marsh into an open-water lagoon. In addition, coastal marsh plants in microtidal environments like our study area typically have elevation ranges of 0.2 to 0.4 m (Silvestri et al., 2005), and, thus would require a sea-level jump larger than these elevation ranges to enable complete drowning of such ecosystems. In light of these observations, we conservatively adopt a value of 0.2 m as the minimum amount of abrupt sea-level rise. Thus, the sea-level jump around 8.2 ka in our study area amounts to 0.20–0.56 m.

Since the catastrophic LAO drainage would perturb the gravitational field and lead to non-uniform changes in sea level (Kendall et al., 2008), sea-level rise observed in the Mississippi Delta would measure only a fraction of the eustatic sea-level rise (this fraction is known as the fingerprint). The fingerprint value could range from 0.2 if the drainage consisted of LAO freshwater only, to 0.4 if the drainage occurred exclusively as rapidly disintegrating ice over Hudson Bay and Hudson Strait (Kendall et al., 2008). As the contribution from disintegrating ice has been proposed to be relatively small (Clarke et al., 2009), we assume a fingerprint value of 0.25 for the final LAO drainage. The observed sea-level rise of 0.20 to 0.56 m at the Mississippi Delta would then correspond to 0.8 m to 2.2 m of eustatic sea-level rise associated with the final LAO outburst (equivalent to ~3 to 8 x 10^4 m^3), exceeding previous estimates for the final LAO drainage (e.g., Barber et al., 1999; Leverington et al., 2002; Törnqvist et al., 2004a). Given the uncertainties in the position of the ice margin of the retreating Laurentide Ice Sheet, the LAO volume may have been larger than the reconstructed 0.45 m sea-level equivalent (SLE) (Leverington et al., 2002) but it is conceivable that the freshwater flux included some Laurentide Ice Sheet melt, likely including icebergs. Therefore, the volume estimate provided here most likely includes both the LAO drainage and discharged icebergs. The relative proportion of these two components, however, is difficult to determine.

A recent study in the Rhine–Meuse Delta inferred a sea-level jump of ~3 ± 1.25 m at 8.54–8.2 ka (Hijma and Cohen, 2010). Although the ~3 ± 1.25 m SLE is larger than our estimate of 0.8 to 2.2 m SLE, we note that our estimate is exclusively associated with the final stage of LAO drainage, while the ~3 ± 1.25 m SLE may well capture multiple pulses of LAO drainage (Hijma and Cohen, 2010). Thus, the two records can potentially be reconciled. Nevertheless, it is the final pulse of LAO drainage that triggered the widespread surface ocean freshening and cooling (Ellison et al., 2006), corresponding to the 8.2 ka climate event as seen in most terrestrial records.

5.4. Implications for abrupt climate change

It has long been postulated that freshwater drainage can trigger abrupt climate events, but large dating uncertainties have prevented causal links from being convincingly established. Our new chronology for the final LAO drainage of 8.18 to 8.31 ka is distinguishable from the timing of the onset of the 8.2 ka event at 8.15 to 8.25 ka (Cheng et al., 2009; Kobashi et al., 2007; Thomas et al., 2007) (Fig. 6b). This allows for a near-instantaneous ocean–atmosphere response to freshwater forcing, consistent with model predictions (LeGrande et al., 2006; Wiersma and Renssen, 2006). Therefore, our study provides independent chronologic evidence for the hypothesized causal link between the final LAO drainage and the 8.2 ka climate event, and currently constitutes a rare firmly established example of a major abrupt climate change that can be tied directly to a well-identified source of freshwater forcing. The vigorous, ongoing debate regarding such a causal link for other abrupt climate events such as the Younger Dryas (e.g., Broecker et al., 2010; Carlson et al., 2007; Firestone et al., 2007; Lowell et al., 2009) highlights the significance of independent age models for both cause and effect with century-scale or better time resolution.

Finally, the new evidence presented here cannot only inform our understanding of the sensitivity of the MOC to freshwater forcing, but also help improve the accuracy of predictive climate models in the context of future increased ice melt as a result of global warming. Given that the freshwater volume that triggered the 8.2 ka climate event likely amounted to more than 0.8 m of near-instantaneous eustatic sea-level rise, our findings lend support to the notion (Meehl et al., 2007) that abrupt cooling due to global warming in the next century is relatively unlikely.

6. Conclusions

We present a high-resolution early Holocene sea-level record from the Mississippi Delta that documents a distinct sea-level jump, marked by a characteristic stratigraphic succession that is corroborated by paleoenvironmental reconstruction. The 0.20–0.56 m local sea-level jump occurred within the 8.18 to 8.31 ka (2σ) time window and is attributed to the final drainage of proglacial Lake Agassiz–Oijibway (LAO). Since the timing of the sea-level jump is indistinguishable from the onset of the 8.2 ka climate event, this study provides compelling evidence for a nearly immediate ocean–atmosphere response to the freshwater perturbation.

In addition, the total inferred eustatic sea-level rise at 8.2 ka (after correction for gravitational effects) amounts to 0.8 to 2.2 m, considerably higher than previous estimates for the final stage of LAO drainage. The new constraints on the timing and amount of final LAO drainage provide additional insight into the sensitivity of MOC to freshwater perturbation, a crucial step toward understanding abrupt climate change. For example, our findings support the notion that abrupt cooling due to global warming in the next century is relatively unlikely.

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