T. C. Moore Jr
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J. C. G. Walker
1

D. K. Rea
1

C. F. M. Lewis
2

L. C. K. Shane
3

See next page for additional authors

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Younger Dryas interval and outflow from the Laurentide ice sheet

T. C. Moore, Jr.,1 J. C. G. Walker,1 D. K. Rea,1 C. F. M. Lewis,2 L. C. K. Shane,3 and A. J. Smith4

Abstract. A boxmodel of the Great Lakes is used to estimate meltwater flow into the North Atlantic between 8000 and 14,000 calendar years B.P. Controls on the model include the oxygen isotopic composition of meltwaters and lake waters as measured in the shells of ostracodes. Outflow rates are highest when oxygen isotopic values of the lake waters are most negative, denoting a maximum glacial meltwater component. Flow rates reach maximum values before the onset of the Younger Dryas and after it ends. These maxima appear to be correlative with the major meltwater pulses MWP 1A and 1B. Although the resumption of North Atlantic Deep Water formation may be tied to the reduction in ice sheet melting, neither the onset nor the end of the Younger Dryas, as recorded in the Greenland Ice Sheet Project (GISP2) records, appear tied to maxima in meltwater outflow from the Laurentide ice sheet.

1. Introduction

At the end of the Last Glacial Maximum, the Northern Hemisphere warmed through a series of climatic oscillations. The initial warming, labeled the Bolling-Allerød interval in Scandinavia and northern Europe, was marked by cooler intervals such as the Older Dryas and the inter-Allerød cool pulse (IACP). It was the very pronounced cooling that occurred at the end of the Allerød, the Younger Dryas interval, that has served to divide the deglaciation cleanly into an early and late phase.

As the Northern Hemisphere ice sheets gradually melted back from their maximum extent sometime near 20 ka, meltwaters were funneled either directly or indirectly into the North Atlantic Ocean. It has been proposed [Rooth, 1982; Broecker et al., 1988] that large influxes of meltwaters from these ice sheets may have placed a stabilizing freshwater "lid" on the North Atlantic Ocean and prevented the formation of North Atlantic Deep Water (NADW). Ocean circulation models [e.g., Manabe and Stouffer, 1997] have indicated that freshwater outflows from North America and Europe into the North Atlantic of the order 0.1 sverdups (1 Sv = 1 x 10^6 m^3/sec) for 500 years would be sufficient to prevent vertical turnover and retard the formation of NADW. Some have proposed that the distinct Younger Dryas climatic cooling between ~10 and 11 radiocarbon years B.P. [Mangerud et al., 1974] was a direct result of such a rapid melting phase of the ice sheet and a subsequent shutdown of vertical turnover in the North Atlantic. However, detailed records of bottom water chemistry [Boyle and Keigwin, 1982, 1987; Keigwin and Jones, 1989; Keigwin et al., 1991; Sarnthein et al., 1994], and sea level rise [Fairbanks, 1989, 1990], seem to indicate that the process and pattern of circulation in the North Atlantic has had a much more complex history.

A detailed history of ice sheet melting, including volumes of meltwater produced and the timing of outflow, is needed to compare with both the oceanic history of deep water chemistry and the models describing the impact of meltwater influx. Developing such a detailed history has not been an easy task. It has been approached by looking at near-shore oxygen isotopic records that indicate inputs of the isotopically very light waters derived from the melting ice sheets [e.g., Kennett et al., 1985; Keigwin and Jones, 1995; Bodén et al., 1997]. These records serve to establish timing (within the accuracy of radiocarbon dating); however, there are relatively few such records, and without good spatial coverage it is difficult to develop accurate estimates of the volumes of outflow.

On land, careful mapping and dating of moraines have allowed researchers to establish the retreat history of the Laurentide ice sheet [Mayewski et al., 1981; Dyke and Prest, 1987] and assuming a relatively stable topographic profile of the ice sheet, to make an estimate of the volume of water released as a function of time [Teller, 1990 a,b; 1995; Licciardi et al., 1998, 1999]. While this approach has been very useful, it has a few weaknesses: (1) the meltback rate is averaged over relatively long intervals of time (200 to 500 years); (2) the assumption of a stable dynamic profile for the ice sheet may be incorrect in certain intervals of time (200 to 500 years); (3) published estimates of outflow do not take into account evaporation and evapotranspiration of meltwaters as they flowed through proglacial lake systems. Evaporation may have greatly affected both the volume and isotopic signature of the fresh waters actually reaching the ocean.

Much of the history of the Laurentide melting post-13,000 calendar years B.P. is contained in the basinal sediments of the Great Lakes basins and other proglacial lakes that surrounded the Laurentide ice sheet [Lewis and Anderson, 1992, 1989; Colman et
2. Methods

The large lake level oscillations (~140 m) in the Great Lakes basins has been the subject of numerous studies that integrated the dating of past shore lines and the draining and flooding of small basins surrounding the Great Lakes with an evaluation of the isostatic rebound associated with the withdrawal of the Laurentide ice sheet (summarized by Lewis and Anderson, 1989; Lewis et al., 1990, 1994; Rea et al., 1994 a,b; Moore et al., 1994; Dettman et al., 1995; Safarudin and Moore, 1999). Ostracode shells from these lake sediments provide an excellent record of the isotopic composition of the lake waters. Through a combination of oxygen isotope studies and radiocarbon dating of lake level history and crustal rebound, we can deduce (1) the flow path of Laurentide meltwaters as they fed through the large proglacial lakes (Lakes Agassiz and Barlow-Ojibway), the Great Lakes, and into the Atlantic [Lewis and Anderson, 1989; Lewis et al., 1994; Colman et al., 1994; Rea et al., 1994b](Figure 1), (2) the timing of melting events [Rea et al., 1994a; Godsey et al., 1999], and (3) the volume of meltwaters derived from the ice sheet [Moore et al., 1995; Walker et al., 1998]. In recent months, additional isotopic data from Lake Winnipeg (glacial Lake Agassiz) [Lewis et al., 1998; Rodrigues and Lewis, 1999] and a careful evaluation of past lake volumes and areas [Gareau et al., 1998, 1999] have allowed us to use the inked box model developed by Walker et al. [1998] in a new approach to estimating the history of freshwater flow into the North Atlantic. In this paper we describe the nature of this model, the data that constrain it, and the estimated timing and volume of freshwater flow into the North Atlantic from about 14,000 to 8,000 calendar years B.P.

2.2. Oxygen Isotope Data

Previously published oxygen isotope data from the Great Lakes basins [Colman et al., 1990, 1994; Lewis and Anderson, 1992; Rea et al., 1994a; Dettman et al., 1995] show a range in values of 15-20‰. This very large isotopic signal is believed to reflect varying proportions of isotopically very negative meltwaters in the basins. Oxygen isotope measurements were made primarily on shells of Candona subtriangulata, an ostracode species that thrives in cold, clear lake waters and precipitates shells in near isotopic equilibrium with the water in which it lives [Dettman et al., 1995]. Isotopic values from calcitic ostracode valves were reported relative to Vienna Peedee Belemnite (VPDB) and converted to values relative to VSMOW by adding -3.5‰, a value determined from the $\delta^{18}$O systematics of the modern Great Lakes [Rea et al., 1994a]. Down-core oxygen isotope records are available from the Michigan, Huron, and Erie basins [Colman et al., 1990, 1994; Lewis and Anderson, 1992; Rea et al., 1994a,b; Dettman et al., 1995; Godsey et al., 1999] (Figures 1, 3, and 4 and appendix 2). Of these, the records from the Huron basin are key, for all meltwaters flowed through this basin (and the contiguous North Channel and Georgian Bay) during the time interval studied.

2.3. Timescales

Corels from the Great Lakes basins were correlated using seismic stratigraphy, lithostratigraphy, grain size, and magnetic
Figure 3. Oxygen Isotope data used as input to and checks on the Great Lakes flow model. (a) Data from proglacial Lake Agassiz used as input to the flow model (references as noted). The data from Rodrigues and Lewis [1999] and from Last et al. [1994] were measured on ostracode valves and reported relative to VPDB. As with the Great Lakes ostracode oxygen isotope data [Rea et al., 1994a], we have added -3.5‰ to these values to convert them to values relative to VSMOW. The data from Remenda et al. [1994] were measured on pore waters and reported relative to SMOW. (b) Data from the basins surrounding Lake Huron used as checks on the drainage pathways and the isotopic composition of Huron waters (references as noted).

susceptibility data [Rea et al., 1994a; Moore et al., 1994]. Selecting coring locations within an interpreted network of seismic reflection data has allowed us to sample most of the ~60 m thick sedimentary section in the Huron and surrounding basins using our relatively short, 6 m piston corer. For that part of the core records less than ~10,000 radiocarbon years B.P., timescales were established using accelerator mass spectrometry (AMS) radiocarbon dates of either shell carbonate or organic carbon samples from the cores studied (Figure 2). Shell carbonate dates were corrected for hard water effects [Rea and Colman, 1995; Moore et al., 1998]. For the Main Algonquin lake highstand and older intervals, ages of lake level changes are based on dated sequences from within the Great Lakes basins and from surrounding bogs and smaller lake basins (summarized by Lewis and Anderson [1989] and Lewis et al., [1994]; Figure 2 and appendix 1). We have applied these dates to the major lithologic changes seen in our Great Lakes cores [Rea et al., 1994a] and to pollen data from these cores [Lewis and Anderson, 1989, 1992; Anderson and Lewis, 1992]. Ages of samples within the varved intervals are based on the varve counts relative to one of these dated points.

All radiocarbon dates have been converted to calendar years using the CALIB program (4.1.2) of Stuiver et al. [1998] [see also Stuiver and Reimer, 1993] using the atmospheric decadal data set for lake core data and the marine decadal data set (with associated reservoir corrections contained in the program) for marine core data. Ages for individual core samples were interpolated between these converted radiocarbon ages using either linear interpolation or smooth curves (where second-order polynomials fit the data with no reversals in slope and with r values >0.99). Sample spacing for δ18O measurements within individual cores ranged from 5 to ~100 years, with occasional gaps of several hundred years where no ostracode shells were preserved.

Having established a consistent timescale in calendar years for...
Figure 4. Correlation of isotope data. (a) Correlation of M17 [Rea et al., 1994b] and 94800-1 (appendix 2) to the Long Point core [Lewis and Anderson, 1992]. (b) Correlation of core 94800-1 (open diamonds) with cores 3P (solid diamonds), 22P (inverted triangles) and 37P (triangles). Correlations between the latter three cores are based on lithostratigraphic components [Rea et al., 1994a]. Core 3P is from Georgian Bay and has isotopic values somewhat more negative than the main part of the Huron basin. Note particularly the very negative values during the middle Mattawa highstand, indicating continued flow of meltwaters into this relatively shallow bay. (c) Average oxygen isotope data for all Huron basin cores (each core with a different symbol) shown versus Huron basin lake levels (from Figure 2). Solid circles = M17 [Rea et al., 1994b]. Plus signs (bottom) and small solid diamonds (top) represent core 94800-1 (appendix 2). Small open diamonds represent core LH91-3P [Rea et al., 1994a] (appendix 2). Solid triangles represent core LH91-22P [Rea et al., 1994a]. Open triangles represent core LH91-37P [Rea et al., 1994a]. Solid diamonds represent core LH95-34P (appendix 2).
Evaporation rates were generally held constant at modern values. Drainage basins [e.g., Shane and Anderson, 1993; Schwalb et al., 1995] clearly indicate drier climates in at least some parts of the Great Lakes basin into 50-year intervals in order to begin the modeling effort. Not every 50-year interval of time was represented in our data set. For many intervals, there is only one data point; however, we feel that for the purposes of modeling this 50-year grouping achieves the goal of high resolution without attempting to model every data point individually. Where two or more δ18O data points were present in a given time interval, both the average and the most negative values were modeled.

2.4. Model

The model used to estimate outflow rates in this study is a linked box model as described by Walker [1991], in which each major Great Lake basin is represented by a box. The volume of each box is equal to the lake basin volume. Inputs to the model are listed in Table 1, along with the data sources. The lake areas are important in determining the amount of evaporation from the lake surface (expressed in the model in units of cm3/yr). Lake volumes are critical to estimating the residence time of the waters and the downstream impact of their flow. The watershed area plus the lake area sum to nearly a constant value through time and, given a precipitation rate and evapotranspiration rate (also measured in cm/yr), determine the volume of water received by the lake system from direct rain and snow.

The precipitation rates [Botts and Krushelnicki, 1988] and isotopic composition [Rozanski et al., 1993] were taken as modern values except during those times when pollen studies clearly indicate drier climates in at least some parts of the drainage basins [e.g., Shane and Anderson, 1993; Schwalb et al., 1995]. The isotopic composition of the rain may have varied within the time interval studied; however, these changes are assumed to have been relatively small (1 to 2%) compared to the isotopic impact of glacial meltwater (approximately 25%). Evaporation rates were generally held constant at modern values. During the Main Mattawa interval (Figure 2) evaporation rates were raised slightly to match the relatively heavy isotopic composition of waters in the Michigan basin (Figure 3b). Similarly, the relatively light isotopic compositions of Michigan basin waters during lowstand times when no meltwaters flowed directly into the basin (e.g., late Stanley time; see Figure 2) were matched by making the rainfall slightly more negative (up to 11.5%).

Evapotranspiration rates were altered slightly to conform with the types of vegetation present in the drainage basins [Zak et al., 1994; Botts and Krushelnicki, 1988; Greenland, 1987; Rosenzweig, 1968] as indicated by pollen studies [Shane, 1987; Shane and Anderson, 1993, Schwalb et al., 1995]; however, these changes were small (15%) and did not greatly affect the modeled outflow rates insensitivity tests. The model assumes there is no fractionation of the oxygen isotopes during transpiration.

Drainage pathways and the isotopic composition of meltwaters are also taken from previous studies [Lewis and Anderson, 1989; Last et al., 1994; Remenda et al., 1994; Lewis et al., 1994; Rea et al., 1994b; Dettman et al., 1995; Lewis et al., 1998; Rodrigues and Lewis, 1999; Figures 1 and 3a). It was assumed that the isotopic composition of Lake Agassiz waters derived from these studies was typical of the waters flowing into the Great Lakes either directly from the ice sheet or indirectly through lake basins that were more proximal to the ice sheet. In any given model run, all basins on which the ice sheet fronted were assumed to receive meltwaters directly. As the ice sheet retreated farther to the north, only the northernmost basins (Superior, and at times, northern Huron) received meltwaters directly.

In each model run the amount of meltwater discharged into the lake system was varied until the isotopic composition of the lake waters calculated by the model matched the isotopic values determined from ostracode valves in sediment cores from the lake basins (Figures 4, 5). The amount of mixing between Lake Michigan and Lake Huron could be partitioned to flow through Erie, Ontario, and/or through the Ottawa drainage to the St. Lawrence valley. Model runs were carried out for 10 different reconstructions of lake volumes and areas (Table 21) [Gareau et al., 1998, 1999]. Within each modeled time interval, runs were

| Table 1. Sources of Data Used as Input to the Model and as Control for Meltwater Discharge and Flow Through the Great Lakes Drainage System |
|------------------|------------------|
| **Sources**      | **Input Data**   |
| Lake area        | Gareau et al. [1998, 1999] |
| Lake volume      | Gareau et al. [1998, 1999] |
| Drainage (river) area | Gareau et al.[1998, 1999] |
| Precipitation rate | Botts and Krushelnicki [1988], Shane and Anderson, [1993], and Schwalb et al., [1995] |
| Evaporation rate | Botts and Krushelnicki [1988], Shane and Anderson, [1993], and Schwalb et al., [1995] |
| Evapotranspiration rate | Rozanski, et al.[1993] |
| Rain Isotopic composition | Lewis and Anderson [1989] and Rea et al. [1994b] |
| Meltwater isotopic composition | varied to match isotope measurements |
| Meltwater supply rate: | 
| Lake Michigan | Colman et al. [1990, 1994] and Appendix 2 |
| Lake Huron, Georgian Bay | Dettman et al.[1995], Rea et al. [1994a,b], and Appendix2 |
| Lake Erie | Lewis and Anderson [1992] |

Supporting Tables 2 and 3 and Appendix 2 are available on diskette or via Anonymous FTP from kosmos.agu.org, directory APPEND (Username=anonymous, Password=guest). Diskette may be ordered from AGU, 2000 Florida Ave, N.W., Washington DC 20009 or by phone at 800-966-2481; $15.00. Payment must accompany order.
conducted to match the minimum (most negative) and average values of lake water oxygen isotopic composition (as indicated by ostracode measurements; see Figure 5). Output from the model includes the residence time of waters within each basin and the outflow rate from the lake system (Table 3).

It is reasonable to assume that the isotopic composition of the meltwaters coming into the lakes is always more negative than the lake waters themselves because of evaporation from the lake surface and the mixing of isotopically heavier rain water received directly by the lakes. In model calculations the more negative the oxygen isotopic composition of the meltwater relative to Huron basin water, the less meltwater is needed to match the calculated isotopic composition with the measured isotopic composition of the lake waters. Likewise, if the incoming meltwater is close to the isotopic composition of the lake water, a larger rate of meltwater input is needed to match the Huron basin isotopic values. Given the range of input values used in this study (Figure 3a), we conducted a sensitivity test for a lake lowstand (the early Stanley time interval; Figure 2; ~10,650 calendar years B.P.) which has a fairly negative estimated average isotopic value of waters in the Huron basin of -18.3% (VSMOW or -14.8% VPDB; see Table 3). By varying the isotopic composition of the incoming meltwaters from about -19 to -25% (SMOW from Figure 3), the calculated outflow rate ranged from .016 to .084

Figure 5. All Huron basin oxygen isotope data (plus signs) used in model estimates of outflow. Circles denote minimum values modeled. Triangles show average values modeled. Where only one data point exists in the time intervals modeled, the minimum and "average" values coincide.

Figure 6. Results of a sensitivity test to determine the effect on Great Lakes modeled outflow of varying the difference between the isotopic composition of the entering meltwater and the isotopic composition of Huron basin water. Meltwaters were varied between -19.4 and -25.2% (SMOW) in the model to match Huron basin waters held fixed at a value of -18.3% (SMOW). It is assumed that the inflowing meltwaters are always isotopically more negative than the Huron basin waters (see text).
The outflow estimates of this test, plotted as a function of the difference between the estimated isotopic composition of the meltwaters and that of the lake waters (Figure 6), indicate that the biggest impact on the outflow estimate occurs when the isotopic difference between meltwater input and lake water is smaller than 2 to 3%. The isotopic differences used in our modeled estimates of outflow (Table 3) were generally >2‰ and averaged near 6.7‰. Given the rather scant control on the isotopic composition of Lake Agassiz waters (Figure 3a), we suggest that our estimates of outflow are accurate within a factor of 2; with larger errors possible during times of peak melting and outflow when the meltwaters derived from Lake Agassiz might reach their most negative values. Clearly, the establishment of a more detailed record of the isotopic composition of Lake Agassiz waters and chronostratigraphies relating the Lake Agassiz and Great Lakes records would be useful in future modeling efforts.

The model only accounts for the meltwaters and other drainage that passthrough the Great Lakes system. It does not account for the ice and meltwater that flows directly into the St. Lawrence valley, Labrador Sea, and North Atlantic Ocean. To account for these additional sources of freshwater that fed directly into the North Atlantic, we measured the perimeter of the Laurentide ice sheet as mapped by Mayewski et al., [1981], with reference to the drainage pathways described by Teller [1990a,b] and Licciardi et al. [1999]. No correction was made for drainage into the Arctic Ocean. The perimeter of the eastern and southern parts of the ice sheet was partitioned into three sections: (1) that which drained into the Great Lakes system, (2) that which drained into the St. Lawrence valley, and (3) that which drained into the Labrador Sea and the North Atlantic. We constructed a ratio of the length of the ice sheet perimeter draining into the St. Lawrence valley, the Labrador Sea and the North Atlantic Ocean to the length of perimeter that feeds through the Great Lakes system. After converting the radiocarbon ages of the reconstructed ice sheet perimeters to calendar ages [Stuiver et al., 1998], values for this ratio were interpolated between each of the mapped time intervals to match the modeled time intervals and were multiplied by the outflow rates estimated by the model. This estimated additional outflow was then summed with the modeled outflow from the Great Lakes to arrive at our estimate of total outflow rate. The basic assumption underlying this adjustment is that variation in melting rates derived from our model of the Great Lakes system is representative of the ice sheet as a whole. This is undoubtedly an oversimplified assumption, especially in light of ice streams that must have carried large volumes of ice directly into the Labrador Sea. However, it does provide what is probably a conservative estimate of the total outflow, with pulses of flow tied directly to melt flow through the Great Lakes and constrained by the outline of the ice sheet existing at the time of the modeled interval.

3. Results

3.1. Comparison of Oxygen Isotope Data From the Great Lakes and the GISP2 Ice Core

The age model developed for the Great Lakes records and their relationship to the standard Great Lakes lake level history [Lewis and Anderson, 1989; Lewis et al., 1994] and to the seismic stratigraphy of the lake cores [Moore et al., 1994; Rea et al., 1994b] is presented in Figures 2-4 and discussed in appendix 1. Using this age model, we can compare the record of lake levels and climate in the Great Lakes to the record of the GISP2 ice core [Grootes et al., 1993; Meese et al., 1994; Stuiver et al., 1993; ]

Figure 2]. Given the errors of our radiometric dates (which are rarely less than ±200 years in the earlier part of our record), it is surprising how good the match between the two records for the onset of the Younger Dryas appears. The initial rise in Lake Algonquin (event 1, Figure 2) occurs near the IACP (the very negative event at 13170 calendar years B.P.). Lake isotopes reach their highest values (Figure 5) near the time of the final Allerød warming over Greenland at ~13000 calendar years B.P. (between events 1 and 2 in Figure 2 and at the relatively warm time of the spruce minimum in the Great Lakes region). The rise in lake levels associated with the draining of Lake Agassiz into the Great Lakes (event 2, Figure 2) is close in time to the initial drop to the most negative isotopic values in the GISP2 core associated with the onset of the Younger Dryas cooling and with the cooling in the Great Lakes region (events 2 and 3, Figure 2).

In the Great Lakes, there is warming at ~11,790 calendar years B.P. (a decline in spruce abundance; see event 4, Figure 2). Within the accuracy of the dating this warming is identical to, or slightly precedes, the end of the Younger Dryas in the GISP2 ice core (at 11,600 calendar years B.P.). The major fall in lake levels from the Main Algonquin highstand appears in our cores to have occurred a few hundred years later (event 5, Figure 2) as the Laurentide ice sheet melted back and opened a deeper passage to the east. Following the Younger Dryas step, isotopic values in the GISP2 ice core became slowly more positive with several negative excursions (Figure 2). The accuracy of our timescale does not allow a detailed correlation of these excursions with melting pulses in the Great Lakes region.

Pollen data from the region surrounding the Great Lakes suggest continued relatively warm conditions, with a return to cooler conditions near the beginning of the Main Mattawa highstand, well after the end of the Younger Dryas (Figure 2) [Anderson and Lewis, 1992; Lewis and Anderson, 1989]. Thus, during the younger part of the Main Algonquin highstand and through the ensuing the lowstands of lake levels we have a comparatively warm climate in North America with outpourings of very negative meltwaters feeding through the Great Lakes system. Was this outflow high enough to cap the North Atlantic thermohaline circulation either during or following the Main Algonquin?

3.2. Outflow From the Great Lakes System

Reasonably small variations in rainfall, evaporation, or evapotranspiration do not have a major impact on the outflow rates computed by the model; however, variation in the isotopic composition of the meltwater input can have a large effect on estimated outflow rates (see section 2). It is interesting to note that some of oxygen isotope measurements for Lake Agassiz waters (Figure 3a and associated references) were more positive than some of the measured oxygen isotope values in the Huron basin (Figure 4). It is clear from this that the isotopic composition of the meltwaters must have varied considerably, with the most negative values likely to have occurred when the ice sheet was in retreat.

Model runs (Table 3) were able to approximate closely the oxygen isotopic records of the Great Lakes basins (Figure 5 and Table 3), and the highest flow rates calculated by the model are associated with the most negative oxygen isotope values found in the lake records (Figures 5 and 7). These occurred at times when the ice sheet was melting, deeper pathways were being exposed for drainage to the east, and a comparatively large proportion of the waters passing through the lake system were being derived from the melting of the isotonically very negative glacial ice. The
Figure 7. Estimated outflow (in sverdrups) to the North Atlantic from the Laurentide ice sheet (from Table 3). Solid triangles indicate estimated maximum flow from the Great Lakes system itself. Circles indicate average outflow from the Great Lakes system (often close to, or identical to, maximum values). Diamonds denote estimated average total outflow, and open triangles denote maximum total outflow from the Great Lakes plus the St. Lawrence valley and the entire eastern and southeastern margins of the Laurentide ice sheet (see text). Shaded background indicates Huron basin lake levels (from Figure 2).

highest estimated rates of discharge from the Great Lakes system occurred during times of relatively low lake levels, particularly during the fall from the Main Algonquin highstand and the three Stanley lowstands that follow the Younger Dryas (Figure 7). During the earliest (Kirkfield-Algonquin) pulse of high flow rates most of the meltwater passed directly into St. Lawrence valley or into the North Atlantic without going through the Great Lakes. Thus the peak flow rates 13,650 calendar years B.P. derive primarily from regions outside the Great Lakes (Figure 7). There are scant data from the Main Algonquin interval itself, but these indicate fairly low outflow rates in the early part of the Main Algonquin. Both the dominant gray color of this interval and its lack of carbonate microfossils are similar to modern sediments in the basin [Rea et al., 1994a] and to sediments laid down in the winter season of the post-Main Algonquin varves [Godsey et al., 1999]. They are indicative of locally derived sediments, lower accumulation rates, and little or no input of meltwaters. Thus we suggest that meltwater input through the Great Lakes during this cold interval was relatively low.

By the start of the Main Algonquin the southeastern margin of the ice sheet had withdrawn to the northern side of the St. Lawrence valley [Mayewski et al., 1981]. Here its position remained fairly stable till the end of the Main Algonquin phase and the second pulse of melting. Throughout this time the Laurentide ice sheet margin appears to have remained close to the coast all along the western part of the Labrador Sea [Mayewski et al., 1981; Dyke and Prest, 1987]. The melting phase associated with the end of the Main Algonquin occurred after the end of the Younger Dryas and reached the highest values of meltwater flow-through modeled for the Great Lakes system itself during the lake level fall and in the subsequent Stanley lowstands.

The final pulse of flow during the late Stanley phase is something of an anomaly. The estimates of flow rates (Figure 7) derive from the very negative oxygen isotope values found in the Huron and Georgian Bay basins (Figures 4 and 5) [Rea et al., 1994a, b] from 8500 to 8900 calendar years B.P. This time marks the warming following the end of the Main Mattawa cool interval of lake highstand [Anderson and Lewis, 1992] and just precedes the final collapse of the ice sheet (8000 to 8400 calendar years B.P.) [Barber et al., 1999]. Light oxygen isotopic values are also measured in northern Lake Michigan at this time (core LH95-16P, Figure 1 and appendix 2). When the Laurentide ice sheet collapsed, the proglacial Lakes Agassiz and Barlow-Ojibway and the remaining meltwaters emptied to the northeast through the Hudson straits. We believe that the very negative isotopic values found in the Great Lakes basins represent the final highstand of these very large proglacial lakes, which back flooded into the Great Lakes basins just prior to their release. As in the earlier lake level falls, there are oscillations in both the oxygen isotope composition of shells and sediment grain size that indicate rapid changes in lake levels and meltwater influx that are apparently related to rapid changes in the ice sheet front and the associated drainage pathways. If these very negative isotopic values do result from back flooding, the application of our model to estimate flow rates through the system during this time is inappropriate. The rates estimated serve to emphasize the magnitude of this "last gasp" of the ice sheet, but may not accurately represent flow through the Great Lakes system. This event in the Great Lakes basins also serves to mark the onset of the final melting phase. It predates the final draining of the more northern proglacial lakes (8000 to 8400 calendar years BP) [Barber et al., 1999], the impact of which is likely seen in the GISP2 record at 8250 calendar years B.P. (Figure 2).

We are modeling only that flow which passes through the Great Lakes system and into the Ottawa/St. Lawrence drainage pathway to the North Atlantic Ocean. Also shown in Figure 7 are estimates (see section 2) of outflow that include drainage from the ice sheet that feeds directly into the St. Lawrence valley east of the Great Lakes and all other drainage from the ice sheet that feeds into the Labrador Sea and North Atlantic Ocean. These
estimates assume that the times of melting are identical around the southern and eastern margins of the ice sheet and that the rates of melting in various areas are proportional to the perimeter of the ice sheet. Both the initial Kirkfield-Algonquin pulse of high flow rates and the first pulse of melting following the Main Algonquin highstand exceed the 0.1 Sv indicated by Manabe and Stouffer [1997] as the flow needed to shut off NADW formation (if applied over a large area for a few hundred years). The fact that our modeled surges of flow appear to be of relatively short duration suggests that large freshwater pulses of relatively short duration are the more appropriate scenario in coupled ocean-atmosphere models of deglacial thermohaline circulation [e.g., Manabe and Stouffer, 1995].

The oldest modeled peak outflow predates the IACP and the subsequent peak outflow from the Baltic ice lake and Fennoscandian ice sheet [Bodén et al., 1997; Lehman and Keigwin, 1992; Lehman et al., 1991]. While the eastern side of the Atlantic was receiving meltwater input near the beginning of the Younger Dryas, outflow from the Laurentide ice sheets appears to have been low and relatively stable (Figure 7) [de Vernal et al., 1996]. The continental climate of the central part of North America appears to have paralleled that recorded in the GISP2 ice sheet through the Younger Dryas. Following the Younger Dryas, the Laurentide ice sheet started the second surge of meltwaters into the western North Atlantic in a series of several (3 to 5) peaks of outflow and lake level lowering that spanned a period of a few hundred years (from the end of the Main Algonquin phase till the end of the middle Stanley). The question remains, Is there evidence that there was enough freshwater outflow to impact the vertical circulation in the North Atlantic?

### 3.3. Laurentide Outflow and North Atlantic Bottom Water Chemistry

The highest-resolution records of bottom water chemistry in the North Atlantic come from the Bermuda Rise region [Boyle and Keigwin 1982, 1987; Keigwin et al., 1991; Keigwin and Jones, 1989, 1995]. In two cores from this area (KNR31-GPC5 at 4583 m water depth and EN120-GGC1 at 4450 m water depth), both $\delta^{13}$C and Cd/Ca ratios in benthic foraminifera were used as proxies for the chemical age of bottom waters. When values of $\delta^{13}$C were comparatively low (more negative) and Cd/Ca ratios were relatively high, it is believed that North Atlantic Deep Waters did not reach to the 4000 m depths of these sites and were replaced by deep waters that derived from the Southern Hemisphere.

Conversion of the radiocarbon dates [Keigwin and Jones, 1989, 1995] from these cores to calendar years and an application of the correlation between the cores [Keigwin et al., 1991] allows a comparison of the proxies of NADW with the GISP2 oxygen isotope records [Grootes et al., 1993; Meese et al., 1994; Stuiver et al., 1995; Figure 8] and with the outflow history of the Great Lakes (Figure 9). Comparison between the GISP2 data and the carbon isotope data from the Bermuda rise cores shows a marked shift to more positive values (greater NADW) within the Allerød (~14,000 calendar years B.P.), followed by a spike of very negative values occurring coincident with the IACP just prior to the beginning of the Younger Dryas (Figure 8). Carbon isotopic values remained relatively positive in the older part of the Younger Dryas and then experienced a gradual return to more negative values, extending almost 1000 years beyond the end of the Younger Dryas before shifting to the more positive values seen in recent times (Figures 8 and 9).

When compared to the estimated total outflow rate (Figure 9), we see no strong correspondence between the $\delta^{13}$C record and
maximum in outflow rate within the Kirkfield Algonquin (at ~13600 calendar years B.P.). The age of high estimated outflow rates is, however, within the range of the meltwater pulse MWP IA seen in the Barbados sea level record [Fairbanks, 1989, 1990]. During the Younger Dryas, the gradual shift to more negative values of $\delta^{13}$C (less NADW) occurs within an interval of no data in the Great Lakes record. The overall drop in lake levels and outpouring of water that mark the end of the Main Algonquin phase matches the meltwater pulse (MWP IB) defined in the Barbados sea level curve (Figure 9) [Fairbanks, 1989, 1990].

This outflow maximum overlaps the most negative $\delta^{13}$C values in core EN120-GGC1, and its subsequent decrease generally tracks the increasing (more positive) $\delta^{13}$C and falling Cd/Ca values that extend beyond the end of the Younger Dryas interval in the GISP2 ice core data (Figure 9a). The solid vertical lines indicate the mean ages of the meltwater pulses (MWP IA and IB) within the vertical shaded bars (1 s variation around the mean) as defined in the Barbados sea level curve [Fairbanks, 1990].

Although the data are very sparse, there is no indication of weakening of NADW formation (Figures 8 and 9) during the final pulse of meltwater outflow indicated in the GISP2 record at 8250 calendar years B.P. Throughout the entire age range of the possible meltwater impact (8900 to 8500 calendar years in this study and 8400 to 8000 calendar years B.P. in the work by Barber...
4. Discussion

The isotopic data summarized here, together with the previously published pollen and lithologic data, allow us to piece together a nearly complete isotopic history of waters in the Great Lakes. One of the surprising results of our earlier studies was that the most isotopically negative waters found in the lakes occurred at times when the lake levels were at their lowest (Figure 4c). The only plausible explanation is that the rapid draining of the lakes and the dominance of isotopically very negative meltwater in the lake basins had a common cause: the rapid melt back of the lakes and the dominance of isotopically very negative meltwater in the lake basins had a common cause: the rapid melt back of the lake basins. The apparent oscillations in lake levels during these overall falls are also consistent with an unstable ice sheet margin that alternately melts back and surges forward. The results of the model runs tend to support this explanation. The meltwater supply and the overall outflow rates are generally higher during the Stanley lowstands than during the Main Algonquin and Matawa highstands.

There may be a substantial error in our estimates of outflow rates from the Great Lakes system associated with the rather incomplete data on the isotopic composition of waters feeding Lake Huron from the more northern glaciolacustrine lakes. However, our sensitivity tests indicate that this error is likely to be within a factor of ±2 of the estimated outflow rate. There is undoubtedly even greater uncertainty in our extrapolation of outflow rates through the Great Lakes basins to the entirety of the eastern and southern parts of the ice sheet, particularly in the older part of the record. Surges of the ice streams that emptied into the far North Atlantic and Labrador Sea could have greatly exceeded our more conservative outflow estimates based on a simple perimeter length relationship. Given these uncertainties, it may be surprising that our estimated outflow rates do reach the magnitudes apparently required to impact NADW formation by the Manabe and Stouffer [1997] vertical circulation model.

The conversion of both the lake and marine records to the calendar year timescale has been critical to our comparison of these records with that of the GISP2 ice core. We assume that the timescale for the GISP2 ice core is accurate to within a few decades. The conversion of radiocarbon dates to calendar years results in a wider 1-s range of age estimates than the associated AMS counting errors because of changes in the radiocarbon composition of the atmosphere with time (Figures 8 and 9). Given these unavoidable errors, the smoothly varying age versus depth curve of the critical marine core (KNR31-GPC5) [Keigwin and Jones, 1989; Keigwin et al., 1991] argues for a robust timescale at this drift deposit site. Although the isotopic records from the northeastern North Atlantic are not as detailed, they are associated with an extremely large number of radiocarbon dates that have been carefully correlated and combined in the synthesis study of Sarnthein et al. [1994]. The multiple stratigraphies and radiocarbon ages used to control the Great Lakes cores place these climatic and isotopic records in a well constrained framework (Figures 2 and 9), and the suite of radiocarbon and U/Th dates used by Fairbanks [1990] also constrains the sea level history independently of the other records.

In spite of the substantial errors in the radiocarbon dates the relationships between these four data sets appear, in general, very reasonable. The nearly exact match in the timing of IACP in the GISP2 record and the δ13C spike in the North Atlantic core records, along with the match of the Younger Dryas onset in the GISP2 record and the onset of the cool interval of the Great Lakes are particularly impressive. The age of the end of the Main Algonquin phase and its associated return to a warmer climate in the Great Lakes region is a key stratigraphic point in this study. The weight of the regional evidence presented in previous studies supports an older age for the end of the regional Main Algonquin cooling than we have used in this study. New data [Yu and Eicher, 1998, and this study] support a younger age which more closely matches the end of the Younger Dryas cold interval. In our cores this warming clearly came before the first major drop in lake levels (the Orange seismic sequence boundary in Figure 2) that marks the end of the Main Algonquin phase, a continuing fall in lake levels, and the outpouring of meltwaters. Thus, even without extremely accurate age control, it is clear that the regional warming associated with the end of the Younger Dryas in North America preceded the outpouring of meltwater.

A key element of our study is the comparison of the timing of the outflow surges calculated using our model with the marine proxies for NADW and sea level. In our comparison between the lake and marine sediment records, we have used the highest-resolution data set available from cores taken in the deep North Atlantic. The lack of a large number of such relatively high-resolution marine records combined with the failure to convert the radiocarbon timescales of the marine cores to calendar years for detailed comparisons to the ice core records has led to some apparent disagreements over when and to what degree NADW was shut off during the Younger Dryas [Keigwin et al., 1991; Sarnthein et al., 1994]. Having made these conversions in our study, these apparent disagreements largely disappear. Sarnthein et al. [1994] indicate that the formation of NADW was shut off (did not reach below 1800 m) at the times that the more detailed records of Keigwin et al. [1991] show strong minima in their δ13C records, just prior to the abrupt postglacial warming at -14,700 calendar years B.P., as well as at and just following the end of the Younger Dryas (Figure 8). These findings do not preclude the enhanced formation of Glacial North Atlantic Intermediate Water (GNAIW) at shoaler depths following the shutoff of the deeper NADW [Marchitto et al., 1998].

In between these shutdowns Sarnthein et al. [1994] call for a return of NADW formation in the northeastern Atlantic comparable to that found in the Holocene (their Holocene/interglacial mode). However, the higher-resolution records from Keigwin et al. [1991] (from the deep northwestern Atlantic) show a marked increase in δ13C (more NADW) ~14,000 calendar years B.P. (Figures 8 and 9) that is quickly followed by a very brief NADW shutdown associated with the IACP and then a return to more moderate δ13C values that gradually become more negative (less deep NADW formation) through the Younger Dryas. In these cores the most negative values of δ13C are found at and after the end of the Younger Dryas (in agreement with the evidence of NADW shutdown at this time presented by Sarnthein et al. [1994]) (Figure 8). Within the errors of radiocarbon dating the youngest interval of very negative δ13C values appears to be associated with the maxima in meltwater outflow from the Laurentide ice sheet and the associated MWP 1B [Fairbanks, 1990]. The intermediate values of δ13C seen in the deep northwestern Atlantic cores suggest that the production of NADW was much diminished from ~12,900 (near the beginning of the Younger Dryas) to ~11,000 calendar years B.P. (several hundred years beyond the end of the Younger Dryas). This weakening of NADW formation is associated with enhanced
GNAIW formation, which continued to about 10,000 calendar years BP [Marchitto et al., 1998], the approximate end of the second period of maximum Laurentide meltwater flux (Figure 9). There is no indication of either diminished NADW formation (Figures 8 and 9) or enhanced GNAIW [Marchitto et al., 1998] during the final outpouring of Laurentide meltwaters at 8000 to 8900 calendar years B.P.

Both MWP 1A and 1B [Fairbanks, 1989, 1990] are associated with relatively warm intervals and with high meltwater fluxes from the Laurentide ice sheet. The large meltwater flux associated with MWP 1B may have managed to significantly impede the formation of NADW following the end of the Younger Dryas; however, both the earlier meltwater pulse (MWP 1A) and the final meltwater pulse (8900 to 8000 calendar years B.P.) are associated with δ¹³C values more typical of modern NADW formation as indicated by data from both Keigwin et al. [1991] and Sartheine et al. [1994] (Figure 8 and 9).

5. Conclusions

The correspondence of our outflow records with MWP 1B of Fairbanks [1989, 1990] and the proxy record for NADW formation supports the arguments for a link between the massive outflow of fresh waters and a prolonged suppression of NADW formation [Rooth, 1982; Broecker et al., 1988] following the end of the Younger Dryas. However, the same does not hold true or maxima in Laurentide outflow and MWP 1A and the final meltwater pulse near 8200 calendar years B.P. which are associated with relatively well-ventilated North Atlantic deep waters and a warming of the regional climate. Cooler climates do appear to be generally associated with the diminished formation of NADW. The enhanced formation of GNAIW appears to follow shutoff of NADW formation; thus, to some degree, the subduction of near-surface waters in thermohaline circulation continues throughout the transition from a glaciated to a full interglacial state. However, the magnitude and locus of this deep or intermediate water formation is likely to have shifted with time and with location of frontal regions. Meltwater outflux from the Laurentide ice sheet could have had some small direct or indirect impact on climate in the region surrounding the North Atlantic; however, there is no conclusive and consistent evidence that Laurentide meltwater outflow into the North Atlantic had a very strong climatic impact on ice-free regions around the North Atlantic between 8000 and 15,000 calendar years B.P.

It remains to be explained what led to the abrupt end of the Younger Dryas while the suppression of NADW formation and enhanced GNAIW formation [Marchitto et al., 1998] apparently remained in effect. The comparisons made in this study indicate that even though the Younger Dryas cold was experienced in North America as well as in Greenland and Europe, neither the start nor the end of the Younger Dryas interval was related to increased outflow from the Laurentide ice sheet. This relationship was first suggested by the sea level study of [1989, 1990]. The melting of the Laurentide ice sheet is tied to warmer intervals and thus is a response to warming rather than a strong and consistent mechanism for cooling.

Appendix 1: Development of the Age Model for Great Lakes Cores

Workers in North America have generally assumed that the relatively cool interval associated with the Main Algonquin lake level highstand [Lewis and Anderson, 1989] was at least approximately equivalent to the deglacial return to cooler conditions noted in Scandinavian regions and named the Younger Dryas [Mangerud et al., 1974]. We base the age of the initial rise of the Algonquin lake levels on the older of two radiocarbon dates on organic material from Kincardine bog (11,200 ±170) [Karrow et al., 1975] and one date from N. Pentagore River (11,300 ±140) [Anderson, 1979]. These two ages were converted to calendar years [Stuiver et al., 1998] and then averaged to give a date for the earliest rise in Lake Algonquin of 13,160 ±165 calendar years B.P. (Figure 2, event 1). The rapid rise of Lake Algonquin in the Main Algonquin phase is thought to coincide with a major drawdown in the Agassiz basin as it began to flow through the Superior basin (Moorehead phase). This is estimated to have begun ~10,850 ±300 radiocarbon years B.P. (12,905 calendar years B.P.; Figure 2, event 2) [Clayton, 1983; Lewis and Anderson, 1989].

The age of earliest fall from the Main Algonquin highstand is based on radiocarbon dates on gyttja (10,650±265 and 10,800 ±360) [Saarnisto, 1974, 1975]. Converted to calendar years and averaged, these ages give an estimate of 12,810 calendar years B.P. for the start of the fall of lake levels (Figure 2, event 3). There are several lines of evidence that point to the oscillatory nature of lake levels during and following the Main Algonquin highstand, starting with this earliest fall and continuing to the early Stanley lowstand [Lewis and Anderson, 1989; Lewis et al., 1994].

The Long Point core from the Erie basin (Figures 1 and 3b) [Lewis and Anderson, 1992] has its own radiometric age control (using pollen correlations), a record of oxygen isotope variation that spans the Kirkfield–Main Algonquin transition, and a pollen stratigraphy that has been correlated to other dated cores in the area. These data show good correspondence to the pollen [Lewis and Anderson, 1989] and isotope [Rea et al., 1994b] stratigraphy of core M17 in the Huron basin, which allow us to correlate the two cores (Figure 4a). The 2% offset in isotopic values between the lake basins is not surprising given their different size and hydrography. Our age model of the Long Point core (after conversion to calendar years) puts the younger isotope maximum at 13,050 calendar years B.P. and the older maximum at 13,450 calendar years B.P. When these points are aligned with the isotopic data from core M17 [Rea et al., 1994b] the general shape of the two isotope curves is found to be similar, as are the pollen data and lithostratigraphy [Lewis and Anderson, 1989, 1992]. Using primarily the timescale for the Long Point core [Lewis and Anderson, 1992], we see the rapid 3% drop in isotopic values between ~12,900 and 12,720 calendar years B.P. following the youngest isotopic maximum. The older date coincides with the timing of the rapid drawdown of Lake Agassiz dated at about 12,900 calendar years B.P. (Moorehead phase [Clayton, 1983]) at the beginning of the Main Algonquin phase in the lower Great Lakes (Figure 2 event 2) [Lewis and Anderson, 1989]. The end of this interval of fairly negative oxygen isotopes is equivalent in age with the younger date for the early Algonquin highstand seen in Kincardine bog (12,728 calendar years B.P.) [Karrow et al., 1975].

We next tie in core 94800-1 from the Huron basin, located slightly to the north of M17 (Figures 1 and 4a and appendix 2). Both the isotopic data and lithology from the upper part of M17 match the lower part of 94800-1. There is a shift up core from red-gray varves to reddish gray laminations that occurs near the marked minimum in oxygen isotope values in M17 (190 cm) and at the base of the isotopic data in 94800-1 (763 cm). On basis of the Long Point age model this level has an age of 12,820 calendar years B.P. Within the errors involved in dating (~200
radiocarbon years), this is identical to the age given for the oldest fall in the Algonquin lake levels of 12,810 calendar years B.P. (Figure 2, event 3, and above). The M17/Long Point age model takes us up to ~12,600 calendar years B.P. where an oxygen isotope maximum is seen in both cores. Both the M17 and 94800-1 isotopic data end at this point.

The M17 data end in a hiatus; however, the comparatively positive isotopic value at the top of the lower part of the 94800-1 isotopic data (Figure 4a) marks the top of a laminated interval and the base of a rather homogeneous gray layer. This gray layer contains no ostracode shells. The top of the gray layer (at 430.5 cm) is a sharp color and lithologic boundary that is recognized in many of the cores taken from the Huron basin, Georgian Bay, and the North Channel (Rea et al., 1994a). It is coincident with the Orange seismic sequence boundary that is thought to mark lake level fall at the end of the Main Algonquin phase (Moore, et al., 1994; Lewis et al., 1994).

In finishing dated lithostratigraphic point that allows us to better constrain the age of the top of the gray layer comes from the pollen record. In records recovered near the Great Lakes the relative abundance of spruce pollen increases markedly from a minimum near 11,000 radiocarbon years B.P. (13,000 calendar years B.P. [Lewis and Anderson, 1989]) and is associated with regional climatic cooling and rising lake levels of the Main Algonquin. Lewis and Anderson [1989] mark the subsequent warming and shift from spruce-dominated to pine-dominated flora (along with the disappearance of significant percentages of herbs and shrubs) at 10,500 radiocarbon years B.P., just before a major drop in Main Algonquin lake levels (Figure 2, event 4).

However, Yu and Eicher [1998], in a detailed study of two small lakes just east of Lake Ontario, place this warming and pollen shift closer to 10,000 radiocarbon years. To resolve the substantial difference in age for this datum, we have dated this horizon at 10,160 ±100 radiocarbon years B.P. (11,790 calendar years B.P. [L. Shane, unpublished data, 1999]) in an apparently continuous section from a lake from southwestern Michigan (Emmons Lake, Kent County). We use this new date for the spruce-to-pine pollen shift in developing our chronostratigraphy.

In cores from the Huron basin, the spruce-to-pine shift occurs within the homogenous gray layer (i.e., before the end of the Main Algonquin highstand) and is found between 480 and 510 cm in core 94800-1 (L. Shane, unpublished data, 1999). Using this tie point in the core, we estimate the top of the homogeneous gray layer to be ~11,140 calendar years B.P. (Figure 2, event 5).

Core LH91-3P from Georgian Bay also contains the uppermost part of the gray layer at its base and sufficient ostracodes for isotopic measurements down to the gray layer. Although somewhat more negative than values measured in the northern Huron basin, the isotopic record from 3P can be correlated with that from the upper part of core 94800-1 (Figure 4b). In core 94800-1, varves are better preserved and can be more easily counted than in 3P [Godsey et al., 1999]; thus, by lithologic correlation, correlation of the isotopic data, and varve counts, the age of the records of 3P and 94800-1 are extended up core from our best estimate for the age of the top of the homogeneous gray layer. This gives us an estimated age for the very negative isotopic values found at the base of the upper varved sequence in core 94800-1 and within the isotopic record from 3P of 11,050 calendar years B.P. The interval of thicker varves within core 94800-1 [Godsey et al., 1999] also has a very negative isotopic value, estimated to have an age of 10,920 calendar years B.P. (Figure 2, events 6 and 7). We relate these sharp negative excursions in the oxygen isotope record to melt-back events associated with lake level drawdowns in the long-term fall in lake levels from the Main Algonquin highstand [Godsey et al., 1999].

For the younger time intervals the age scale for the isotopic data (Figures 4b and 4c) is based on radiocarbon dates made directly on the shell material in the cores and on the correlation between cores using lithostratigraphy and magnetic susceptibility [Rea et al., 1994a, b].

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