

A 14,000-year environmental change history revealed by mineral magnetic data from White Lake, New Jersey, USA

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Abstract

The increasing use of magnetic parameters as a proxy for environmental change necessitates the understanding of processes that link magnetic properties of sediments, especially organic-rich lake sediments, and environmental change. To explore the magnetic mineral-paleoenvironment link, we have recovered a 14,000 yr mineral-magnetic record from White Lake, a hardwater lake containing organic-rich sediments in northwestern New Jersey, USA. Pre-14 ka (1 ka = 1000 cal yr BP) sediments are dominated by clays, and the magnetic variations of the clays recorded the deglaciation process, the initial development of the lake, and the stabilization of the watershed. The rapid increase in organic matter and carbonate around 14 ka likely marks the Bølling warming event. The marl sediments were deposited from 14 to 11 ka and displayed a continued decline in magnetic mineral concentration, which indicates increasing marl precipitation in the lake and stabilization of the watershed by vegetation growth. The Holocene gyttja-dominated sediments generally have weak magnetizations. However, yellowish marl layers in the Holocene sediments display strong magnetizations. The strong magnetization of marl layers at ~1.3, 3.0, 4.4, and 6.1 ka probably resulted from oxidation of the precipitated marl sediments near to shore, which were then eroded, transported, and redeposited during periods of low lake levels. This study demonstrates that multiple magnetic parameters can provide a more detailed and robust interpretation of environmental change than lithologic data alone.

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1. Introduction

Changes in environmental conditions can cause variations in the concentration, grain size, and/or composition of magnetic minerals in sediments [1–3]. In recent years, lake sediments have become attractive targets for magnetic measurements in studying past environment and climate changes (e.g. [4–7]), partly because lake sediments can often provide higher temporal resolutions than marine sediments and provide climate records for

continental regions in addressing regional responses to large-scale climate change. Interpretation of magnetic properties of lake sediments often assumes climatically-controlled detrital flux from the watershed into the lake by fluvial or eolian processes. This erosional model can usually provide satisfactory explanations for magnetic properties of sediments deposited during cold periods or in arid regions when the watershed vegetation is sparse and the erosion is intense. The erosional model, however, may encounter difficulties explaining the magnetic properties of sediments deposited during warm periods due to the high content of organic matter that can cause dilution, dissolution, and diagenesis of magnetic minerals.

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Previous studies of glacial–interglacial cycles have shown that sediments deposited during interglacials have much lower magnetic mineral concentrations than those deposited during glacial periods [5,8,9], which have been

interpreted to result from reduction diagenesis (e.g. [8,9]). However, low magnetization during interglacials may still retain important paleoenvironmental information. For instance, Rosenbaum et al. (1996) [10] demonstrated that the low magnetic mineral concentrations of sediments deposited during warm periods could result from low levels of peak runoff. Also, postdepositional diagenesis could represent an indirect response to climate change. For example, the presence of authigenic greigite may indicate a dry climate [11–13].

Despite these findings, the physical and chemical processes that link magnetic mineral properties to environmental change in Holocene organic-rich lake sediments remain largely unknown. Interpretation of magnetic parameters often requires comparison of magnetic properties to other independent climate proxies. Few magnetic mineral-multi-proxy studies have been done on calcareous lake sediments, because their magnetism is typically weak. Calcareous lake sediments, however, are particularly important to paleoclimate studies because they yield oxygen isotope data that are a straightforward and robust proxy for climate (e.g. [14,15]). Studying the magnetic properties of calcareous lake sediments will advance our understanding of the magnetic mineral-paleo-environment link.

Here we present the >14,000 yr mineral-magnetic results of two sediment cores from White Lake, a lake containing organic-rich calcareous sediments in northwestern New Jersey, USA. The magnetic study is part of an ongoing multidisciplinary investigation of lake sediments from the region for reconstructing environmental change. We show that the White Lake sediments record the latest Pleistocene deglaciation, development of the lake and its watershed, as well as lake level fluctuations during the middle and the late Holocene.

1.1. Study site

White Lake is a small hardwater lake (0.26 km² surface area, ~2 km² catchment area) located in northwestern New Jersey (Fig. 1A, B). The lake has a tiny ephemeral inlet at its northeast end and one outlet at its

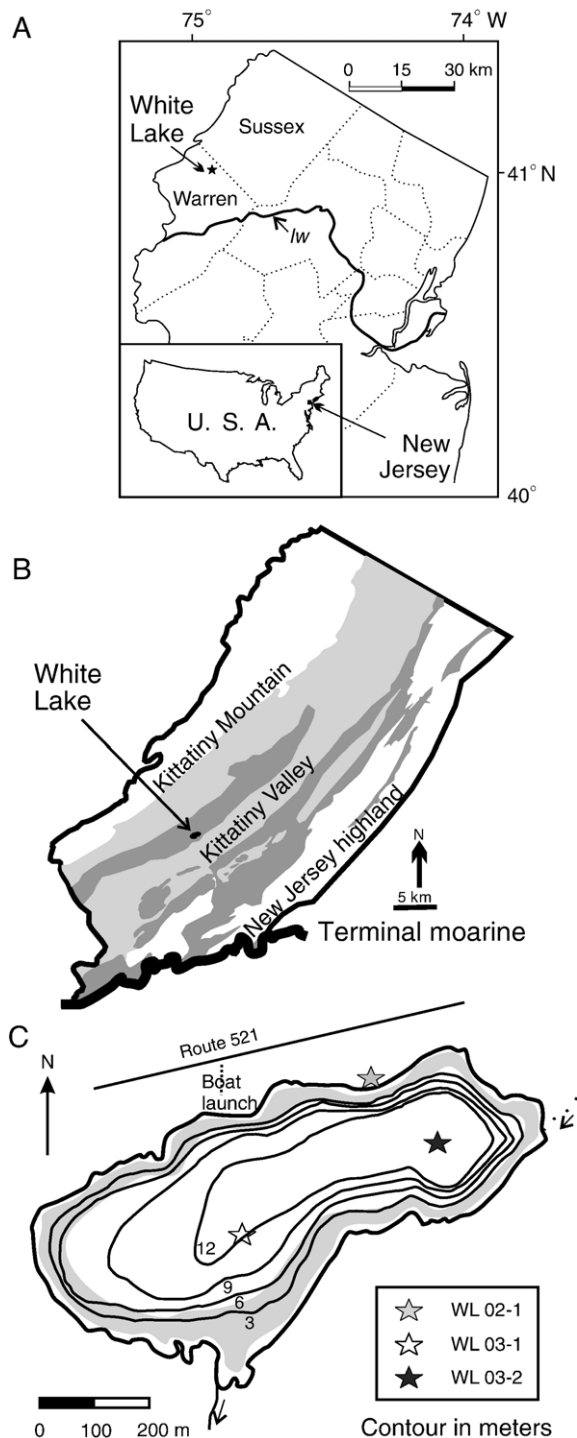


Fig. 1. (A) Simplified map of northern New Jersey showing the location of White Lake and the Late Wisconsinian terminal moraine (bold line marked by *lw*; modified after Witte, 2001). Inset shows the location of the study area in the United States. (B) Physiography and bed rock geologic units of northwestern New Jersey. White = major uplands; Dark gray = carbonate bedrock; Light gray = slate, siltstone, and sandstone (Martinsburg Formation) (modified from Witte, 2001 [16]). (C) Bathymetric map of White Lake showing the locations of the three coring sites. The gray band denotes the marl bench.

south end (Fig. 1C). A marl bench, a band of unconsolidated calcareous deposits, appears along most parts of the shore zone (Fig. 1C). The White Lake basin is believed to have originated in a preglacial karst. However, the lake was not formed until after the Wisconsinan Glaciation when the White Lake basin was dammed by glacial debris at its southeastern outlet, impounding the meltwater as the climate became warm and the glacier front retreated to the north [16,17].

2. Methods

2.1. Field sampling, loss-on-ignition, and radiocarbon dating

Two cores WL03-1 and WL03-2 were recovered from White Lake (Fig. 1) with a Livingstone–Wright piston corer of 5 cm in diameter on February 16 and March 1, 2003, respectively [18]. Another core (WL02-1) was taken from the wetland area on the northern shore for detailed studies of deglacial climate oscillations [19].

Sequential loss-on-ignition (LOI) analysis typically at 4 cm interval (ranging from 2 to 20 cm) was performed to estimate the organic matter (OM) as weight loss at 550 °C and carbonate content of the lake sediments as weight loss at 1000 °C [20]. The balance was considered to be the silicate content. Terrestrial macrofossils at five levels from each core of WL03-1 and WL03-2 were collected for the accelerator mass spectrometry (AMS) ^{14}C dating. The ^{14}C dates were calibrated with the *CALIB* [21] using an IntCal 04 dataset [22].

2.2. Magnetic measurements

Samples for magnetic measurements were collected typically at 4 cm interval in plastic cubic boxes (8 cm³), and all magnetic parameters are normalized by dry mass. Low field magnetic susceptibility was utilized to estimate the concentration of magnetic minerals and was measured with a KLY-3S Kappabridge. Isothermal remanent magnetization (IRM) was measured to estimate magnetic mineral concentrations of grains capable of carrying remanence [23]. The samples were first exposed to a forward field of 1.0 T (samples are considered saturated based on stepwise IRM acquisition data of representative samples, i.e. $\text{SIRM} = \text{IRM}_{1.0}$) using an ASC impulse magnetizer and then a back field of 0.3 T ($\text{IRM}_{-0.3}$). *S*-ratios were computed using $S = \text{IRM}_{-0.3} / \text{SIRM}$ to characterize the relative abundance of low-coercivity magnetic minerals, such as magnetite, and high-coercivity magnetic minerals, such as hematite. The hard IRM (HIRM) was calculated from $\text{HIRM} = (\text{SIRM} - \text{IRM}_{-0.3}) / 2$ to estimate the con-

centration of high-coercivity magnetic minerals, such as hematite and goethite. The anhysteretic remanent magnetizations (ARMs) were acquired in an alternating field with a peak of 100 mT and in a 0.1 mT DC field. The ARM provides a measure of the content of fine-grained ferrimagnetic particles [24,25] and the ARM/SIRM ratio can therefore be used to estimate the proportion of fine-grained ferrimagnetic particles with respect to total amount of magnetic particles for mixtures of ferri- and antiferromagnetic grains. All remanence measurements were conducted with a 3-axis 2G superconducting magnetometer housed in a magnetically shielded room in the Paleomagnetism Laboratory at Lehigh University.

Additional magnetic measurements, including magnetic hysteresis loops as well as high temperature and low-temperature treatments, were conducted for selected samples representative of different compositions from core WL03-2 at the Institute for Rock Magnetism, University of Minnesota. These parameters are used to constrain the interpretation of magnetic mineralogy and the particle size of magnetic grains. Magnetic hysteresis loops and high-temperature thermomagnetic moments were measured with a Vibrating Sample Magnetometer (Princeton Measurements Corp. MicroVSM 3900). Magnetic hysteresis loops were measured with a maximum applied field of 1.0 T to estimate the grain size of magnetic particles [26]. Low-temperature (LT) measurements (<300 K) were performed on a Quantum Design Magnetic Property Measurement System (MPMS) to detect the magnetic transitions that are diagnostic of magnetic minerals such as magnetite [27]. One set of LT experiments was designed to test if there is biogenic magnetite in lake sediments following Moskowitz et al. (1993) [28]. The detailed measurement procedures of LT experiments are described in [29].

Four marl samples from core WL02-1, which presumably have not experienced oxidation, were used to examine whether oxidation of marl sediments could produce magnetic minerals and thus enhance magnetic mineral concentrations. One sample was heated up to 200 °C in air in a 50 μT field for 1 h in the laboratory, emulating a long-term natural oxidation at lower temperatures. The experiment was conducted with an ASC TD-48 thermal demagnetizer that is customized to provide a controlled field and the sample was held in a 2.5 cm diameter quartz cylinder. The other three samples were placed next to a window in the laboratory for five weeks so that samples can be exposed to sunlight and air. The exposure experiments at room temperatures were to deliberately mimic natural oxidation conditions. The magnetic properties of these samples were measured before and after they were oxidized.

3. Results

3.1. The age-depth models

The age-depth models for the two cores were established based on linear interpolation of five calibrated AMS ^{14}C dates (Table 1) from each core and the surface age (2003 AD) (Fig. 2). Multiple materials (e.g., leaf, seed, charcoal), which are listed in the order of their abundance in Table 1, were combined to generate a ^{14}C date at each level because the datable materials are scant in the cores, especially in the upper half of the cores, and individual items (leaves, seeds, etc.) are often too small in abundance to be dated individually. The basal ages of both cores were obtained by extrapolation using the deposition rate of marls.

3.2. Lithology and LOI data

Both cores show a compositional succession from pale grey clay-dominated sediments, through light yellowish marl-dominated deposits, to dark brownish gyttja-dominated sediments. The lower clays appear to contain more silt than the upper clays. The overlying marl sediments are generally dense, coherent, and homogeneous in texture. The remaining gyttja-dominated sediments of the two cores display strikingly different depositional features. In WL03-1, a number of light yellowish marl layers punctuate the dark brownish gyttja sediments; whereas, in WL03-2, only light brownish bands with gradual transitions occurred in the dark brownish gyttja. The yellowish marl layers in WL03-1 comprise mainly coarse-grained materials, and display a heterogeneous texture.

The LOI data of both cores display similar first-order patterns (Fig. 3). The clays contain more than 90% silicate

with a gradual decline starting ~ 15 ka (1 ka = 1000 cal yr BP; Fig. 3). From 14 to 11 ka, the carbonate content shows a slight increase and the silicate content continues to decrease. The gyttja deposited since ~ 11 ka show high OM content, typically ranging from 50% to 85% (Fig. 3). The variation in OM content is mainly due to the relative abundance of carbonate since the silicate content remains consistently low ($\sim 15\%$) throughout the Holocene. As a result, the OM content varies out of phase to the carbonate content: the OM ‘highs’ occurring at carbonate ‘lows’ and vice versa. This pattern of variation is best shown in WL03-1 (Fig. 3A).

3.3. Magnetic stratigraphy

3.3.1. Stage I: pre-14 ka

The clays are characterized by a strong magnetization, multiple types of magnetic mineralogy, and coarse magnetic grain sizes. Overall, the magnetic susceptibility, ARM, SIRM, and HIRM of clays display high values, indicating a high concentration of magnetic minerals (Fig. 3). A rapid increase in magnetic susceptibility, ARM, and SIRM around 18 ka separates the lower clays from the upper clays (Fig. 3). The lower clays show relatively low values of susceptibility and SIRM, suggesting a relatively low magnetic mineral concentration. The low-temperature (LT) data show pronounced magnetic transitions at ~ 35 and ~ 120 K (Fig. 4A). The ~ 120 K transition could indicate the presence of magnetite [27]. The ~ 35 K transition occurs in the LT–SIRM decay curve, but not in the RT–SIRM cooling curve (Fig. 4A), suggesting that other magnetic phases such as siderite, ferrihydrite, and ilmenite (Personal Communication, M. Jackson, 2003) may also be present in the basal clays. The decrease in susceptibility and SIRM in

Table 1
AMS ^{14}C dates from Cores WL03-1 and WL03-2, White Lake, New Jersey

Depth (cm)	Dating materials (in the order of abundance)	C wt (mg)	$\delta^{13}\text{C}$ (‰ vs. VPDB)	AMS lab #	AMS ^{14}C date (\pm SE)	Calibrated age (BP) (mid-point $\pm 2\sigma$ range)
Core WL03-1						
1293–1297	Woody frag., Charcoal, leaf frag.	1.19	–18.6	UCI16922	1875 \pm 20	1821 \pm 54
1391–1394	Seed, charcoal	0.34	–	AA57158	3800 \pm 30	4175 \pm 86
1437–1440	Charcoal, leaf frag., seed	0.70	–	AA57157	6260 \pm 40	7212 \pm 57
1502–1505	Leaf frag., anthers, charcoal	0.84	–28.51	AA56910	9220 \pm 50	10,382 \pm 133
1598–1601	Charcoal, spruce needles	0.48	–	AA57156	11700 \pm 60	13,559 \pm 155
Core WL03-2						
1475–1481	Leaf frag., charcoal, anthers	0.75	–	AA56908	2160 \pm 30	2122 \pm 65
1599–1602	Leaf frag., plant stem, seed, charcoal	0.62	–	AA57155	5750 \pm 35	6547 \pm 94
1700–1702	Leaf frag., seed, larch needles, anthers, charcoal	0.76	–28.08	AA56907	9305 \pm 40	10,491 \pm 111
1743–1745	Birch seed, leaf frag., charcoal	1.06	–17.6	UCI16925	10240 \pm 70	11,974 \pm 268
1802–1805	Charcoal, spruce and larch needles, leaf frag., bark	0.96	–26.96	AA57154	11970 \pm 100	13,841 \pm 216

Note: AMS = accelerator mass spectrometry, wt = weight, frag. = fragments, VPDB = Vienna Pee Dee Belemnite, AA = University of Arizona AMS lab, UCI = University of California, Irvine, Calibration is based on Stuiver and Reimer (1993) [21] and Reimer et al. (2004) [22] using the CALIB 5.0.1.

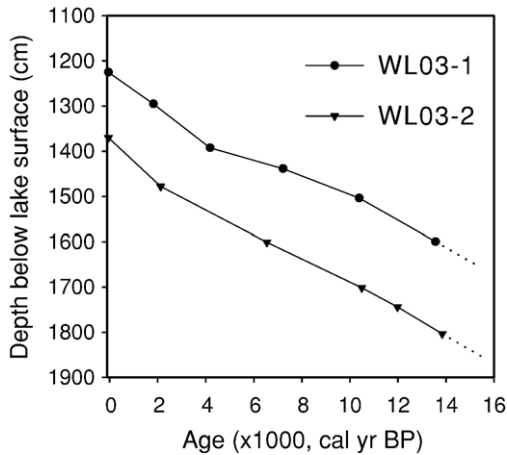


Fig. 2. Age-depth models for cores WL03-1 and WL03-2, derived by linear interpolations of five calibrated ages and surface age (2003 AD) for each core. cal yr BP = calibrated age before present (1950 AD).

the upper clays from ~ 18 to ~ 14 ka suggests a decline in magnetic mineral concentration. The LT data show that the ~ 120 K transition remains pronounced but the ~ 35 K transition becomes greatly subdued (Fig. 4B), indicating an upcore decline in the diversity of magnetic phases. The increased S -ratios (to ~ 0.8) and the decreasing HIRM (Fig. 3) suggests that the high-coercivity magnetic minerals also decline upcore. The generally low S -ratios and high HIRMs of clays (Fig. 3) indicate that the clay-dominated sediments contain abundant high-coercivity minerals. Hysteresis data show that magnetic particles in clays are mainly multidomain (MD) in size (Fig. 5).

3.3.2. Stage II: from 14 to 11 ka

The marl-dominated sediments tend to be transitional in magnetic properties. The magnetic parameters of marl continue the upcore decreasing trend seen in the upper clays, except for the ARM/SIRM ratio (Fig. 3). The ARM/SIRM ratio of marl is clearly higher than that of the clay-dominated sediments, indicating an elevated contribution of fine-grained magnetite particles in the marl-dominated section. This is supported by the high S -ratios (on average, 0.92 and 0.95 for WL03-1 and WL03-2, respectively) (Fig. 3) and the LT data showing a drop at ~ 120 K in the RT-SIRM cooling curve observed by a peak in its 1st derivative curve (Fig. 4C), which probably indicates the presence of magnetite. Hysteresis results indicate that magnetic grains in the marl-dominated sediments are mainly pseudo-single domain (PSD) in size (Fig. 5). Magnetic susceptibility shows a continued upcore decrease with a change from positive to negative values occurring in the marl-dominated section (Fig. 3), indicating that diamagnetic carbonate and OM dominate

the magnetism of the lake sediments by the end of the Pleistocene.

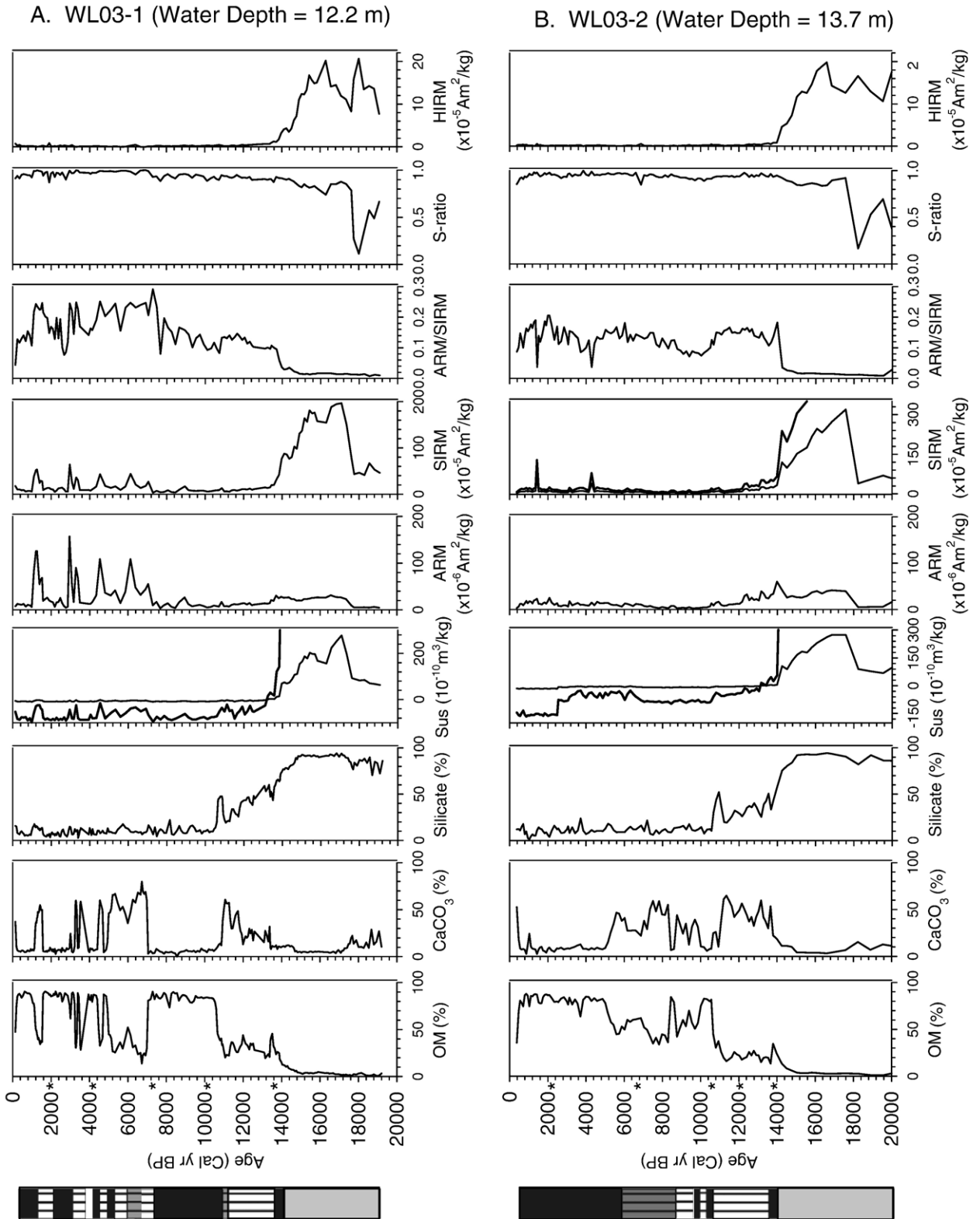
3.3.3. Stage III: after 11 ka

The gyttja show negative magnetic susceptibility values (Fig. 3), indicating low ferromagnetic mineral concentrations. Also, the Holocene samples display high S -ratios (~ 0.95) and low HIRMs, suggesting that a low-coercivity magnetic mineral, most likely magnetite, is the dominant magnetic phase (Fig. 3). The ARM and SIRM parameters of the Holocene sediments from the two cores show distinctively different patterns (Fig. 3). WL03-1 displays an in-phase variation between the ARM and SIRM with peaks occurring in the light yellowish marl layers and troughs occurring in the intervening gyttja layers (Fig. 3A). In contrast, WL03-2 generally shows little variation in the ARM and SIRM, except for two sharp peaks in SIRM (Fig. 3B). The ARM and SIRM of WL03-2 are generally weaker than those of WL03-1 except at the ~ 1.3 and ~ 4.4 ka levels, where SIRM peaks occur that are comparable in magnitudes to those of WL03-1 at similar levels (Fig. 3). These indicate a good intra-lake correlation for the Holocene sediments.

The LT data show a drop in the RT-SIRM cooling curves between 150 and 110 K (Fig. 6A), which may suggest the presence of magnetite particles. The HT thermomagnetic curves show a rapid drop in magnetic moment at ~ 580 °C during heating (Fig. 6B), corroborating the presence of magnetite. The hysteresis data show that the magnetite grains are predominantly PSD in size (Fig. 5). All ARM and SIRM data of the Holocene sediments of WL03-2, except the two SIRM peaks, were compared with the OM content. The SIRMs do not vary with OM content (Fig. 6C), neither do the ARMs (not shown), suggesting that OM does not influence the magnetic signatures. In addition, the gyttja do not appear to contain intact magnetosome chains produced by magnetotactic bacteria based on Moskowitz et al. (1993)'s [28] criteria (Fig. 6A and D). The $\delta FC/\delta ZFC$ ratios of samples from the upper, middle, and lower sections of the gyttja are 1.125, 1.013, and 1.056, respectively. These values are similar to those of the underlying marls (1.045, 1.057), suggesting that magnetite particles in gyttja are likely inorganic in origin [28].

3.4. Magnetic results of the marl oxidation experiments

The magnetic results of the laboratory heated sample and naturally oxidized samples are summarized in Table 2. Both the ARMs and SIRMs of all four samples, particularly the naturally oxidized samples, increased after the experiments, suggesting that the oxidation of marl



sediments may be a feasible process to enhance magnetic mineral concentrations in the light yellowish marl layers of WL03-1.

4. Paleoenvironmental interpretations

4.1. The formation of the lake and its initial development (~18 to ~15 ka)

White Lake is located about 18 km north of the late Wisconsinian Terminal Moraine (Fig. 1). It is believed that the ice retreat initiated ~20 ka in northwestern New Jersey and the front of the recessional ice lobe did not reach the White Lake area until ~18 ka [16]. Therefore, White Lake could have started receiving sediments and recording environmental changes as early as ~18 ka.

The lower clays probably represent glacial lacustrine sediments deposited when the margin of the recessional ice lobe was positioned near White Lake. As the ice melted, the meltwater would have transported sediments into the lake. The lower clays have a low magnetic concentration (Fig. 3), probably because of the limited sediment supply as the White Lake area remained partially blanketed by ice. Also, the lower clays contain both low- and high-coercivity magnetic minerals (Figs. 3 and 4). The diverse magnetic mineralogy is consistent with the fact that ice sheets often acquire sediments from multiple sources along the path of their advance and these sources often contain different magnetic minerals.

The sharp increase in magnetic concentration at the contact between the lower clays and the upper clays (Fig. 3) probably marks a shift of sediment sources from the melting ice to the local watershed. After the ice lobe retreated from the White Lake area, the lake basin was completely exposed to the cold and wet environment, which would promote mechanical weathering and erosion, leading to high sediment yield from the upland area. The abundant erosional debris was transported into the lake by both wind and precipitation (rainfall and snowmelt). The pronounced and gradual decrease in the SIRM upcore (Fig. 3) indicates a reduction in magnetic mineral concentration, thus a decrease in sediment flux into the lake. The decrease in sediment flux likely resulted from

Fig. 3. Lithology, sediment composition, and magnetic results of cores WL03-1 (A) and WL03-2 (B). Lithostratigraphic columns: gray = clays; black = gyttjas; striped = marl; striped with gray background = marly gyttja. The asterisks on the left of the both OM content panels indicate the levels of AMS dates (see Table 1 for AMS ^{14}C dates and calibrated ages). OM = organic matter, sus = susceptibility. The thicker lines of sus were shown as $\times 10$ of the original values and the thicker line of SIRM of WL03-2 was shown as $\times 2$ of the original value. Note that the scales of SIRM for two cores are different.

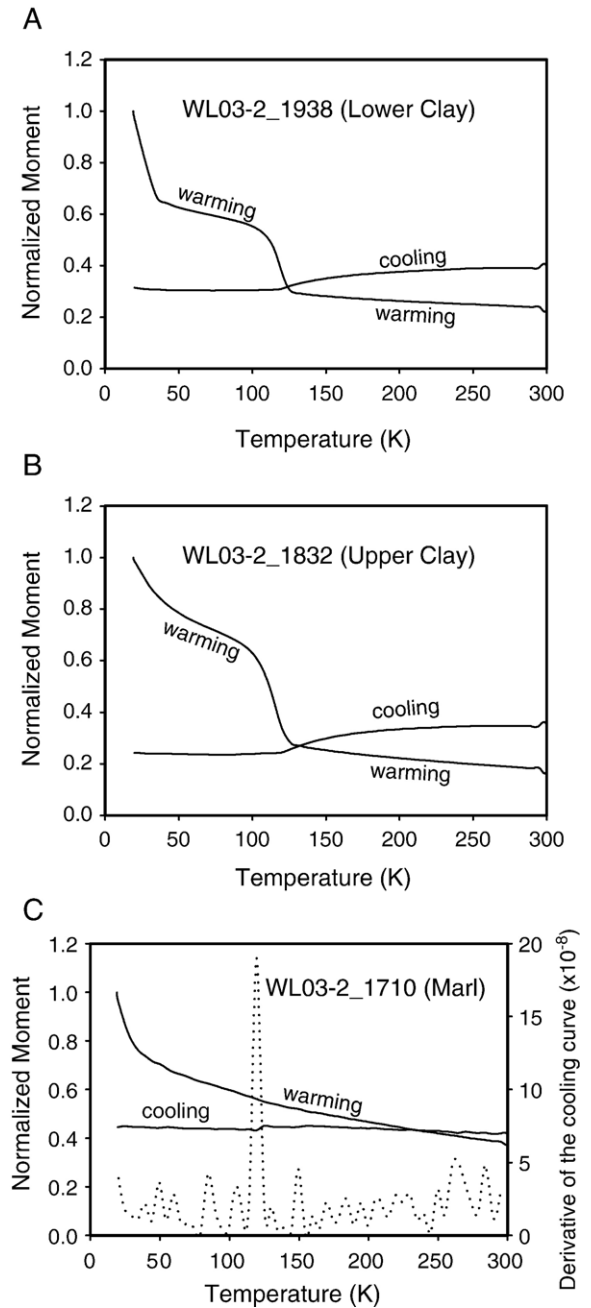


Fig. 4. Low-temperature magnetic properties of the clay-dominated (A, B) and the marl-dominated sediments (C) that deposited during the late Pleistocene. The dotted line in (C) is the 1st derivative of the RT-SIRM cooling curve. RT-SIRM = SIRM acquired at room temperature.

the increased vegetation cover in the upland area, which would reduce the intensity of physical weathering and thus the detritus yield from the upland [10]. As the ice lobe retreated from the White Lake area, herb species would have been established in the watershed and formed a tundra-dominated landscape, as has been suggested by the well-established regional pollen stratigraphy [30]. The

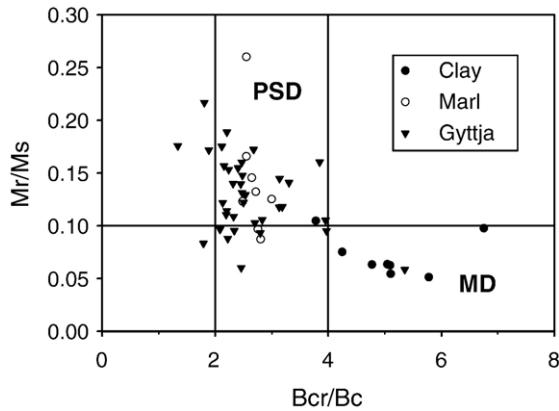


Fig. 5. Hysteresis loop parameters used to estimate grain size of White Lake sediments. Mr = saturation remanence; Ms = saturation magnetization; Bcr = coercivity of remanence; Bc = coercivity. PSD = pseudo-single domain; MD = multidomain.

decreasing HIRM and the increasing S -ratios (Fig. 3) also suggest that the content of high-coercivity magnetic minerals decline upcore in the upper clays. Since the high-

coercivity minerals were likely derived from outside of the watershed by the ice sheet, their contribution to the sediment load would have diminished as the ice lobe retreated from the White Lake area. The clays contain more than 90% silicate (Fig. 3) and generally have coarse-grained (MD) magnetic minerals (Fig. 5), corroborating the interpretation that the clay-dominated sediments represent glacial lacustrine deposits and detritus eroded from the upland.

4.2. The lake and landscape development (~ 15 to ~ 11 ka)

From ~ 15 to 14 ka, the OM and carbonate gradually became dominant constituents of the lake sediments (Fig. 3), suggesting a warming of the climate that appears to be coeval with the Bølling warming documented in the Greenland ice cores [31]. The warming climate made it possible for aquatic vegetation to flourish, facilitating carbonate precipitation in the lake. The warming climate also enhanced the growth of vegetation

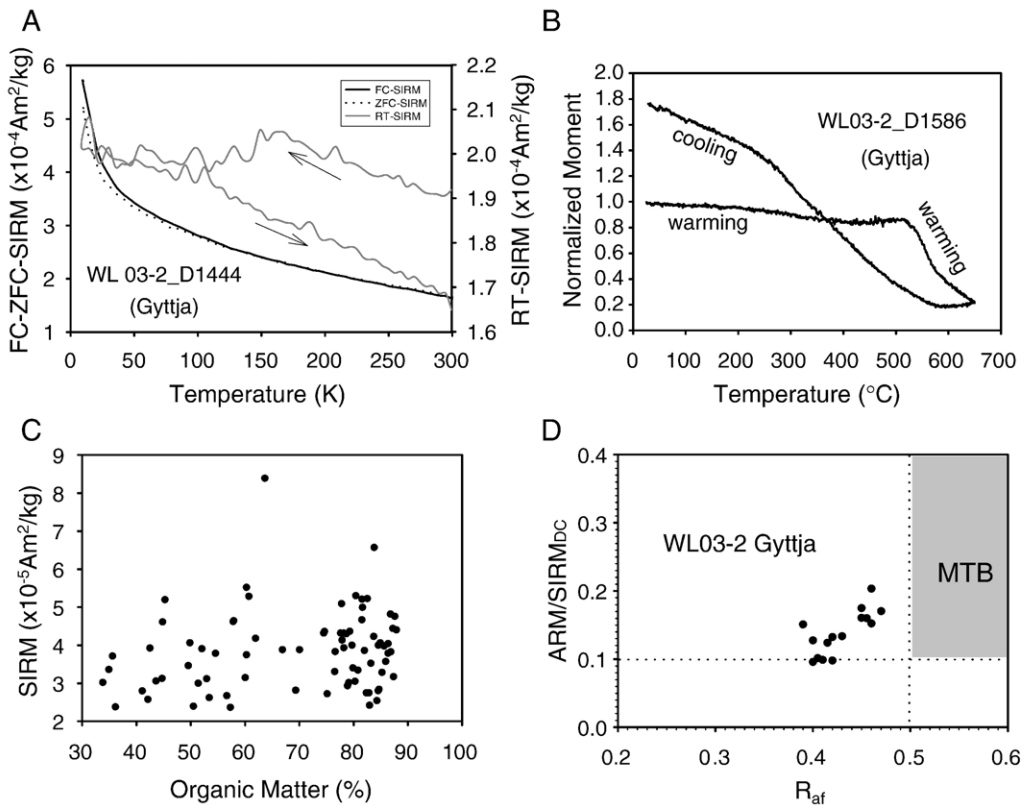


Fig. 6. Magnetic properties of gytija from core WL03-2. The low-temperature (A) and thermomagnetic data (B) suggest the presence of magnetite grains. FC = field cooled; ZFC = zero field cooled; The SIRMs do not seem to correlate with organic matter content (C), and there appears no intact magnetosome chains in gytija (A and D). The shaded area in (D) indicates magnetic parameters diagnostic of magnetotactic bacteria (MTB)-produced magnetite [28]. R_{af} = grain interaction index and $SIRM_{DC}$ = SIRM acquired at a static field at 100 mT. Both were determined following Moskowitz et al. (1993) [28].

Table 2
Laboratory experiments testing the marl oxidation hypothesis

Sample #	Treatment		ARM ($\times 10^{-7}$ Am ² /kg)	SIRM ($\times 10^{-7}$ Am ² /kg)	S ratio	ARM/SIRM
WL-50	Heating at 200 °C	Before	1.19	9.82	0.98	0.12
		After	7.54	252.2	0.94	0.03
		After/before	6.31	25.7		
WL-70	Exposed to sunlight and air for five weeks	Before	1.23	9.73	0.98	0.13
		After	1.55	20.40	0.96	0.08
		After/before	1.25	2.10		
WL-90		Before	0.76	9.75	0.97	0.08
		After	0.84	17.99	0.95	0.05
		After/before	1.11	1.84		
WL-100		Before	0.82	8.47	0.98	0.10
		After	1.29	18.23	0.99	0.07
		After/before	1.56	2.15		

Samples were taken within the top 100 cm of Core WL02-1. ‘before’ and ‘after’ indicate measurements were made before and after treatment. After/before indicates the change in magnetic parameters by calculating the ratio of post-treatment over pre-treatment values.

in the upland, which would continue to stabilize the landscape. Probably, the landscape was largely stabilized by ~ 14 ka when the maximum concentration of both OM and carbonate was reached (Fig. 3) and a dramatic reduction of sediment influx from the upland occurred. The marked break in slope of the SIRM curves at ~ 14 ka (Fig. 3) probably indicates a significant reduction in sediment supply from the upland following the deglacial warming. It is conceivable that detrital limestone debris might have been transported into the lake in the early stage of the lake’s history when climate was cold and vegetation cover was sparse. However, the detrital contribution would be minimal after the watershed has been stabilized since ~ 14 ka as the marl-dominated sediments show a coherent, dense, and homogenous texture.

Magnetic particles in the marl-dominated sediments are mainly PSD (Fig. 5), finer than those (MD) in clays. The shift in grain size is consistent with the development of the stabilizing landscape. As the upland stabilized, fine magnetic grains gradually become the dominant component in the sediment flux transported from the watershed into the lake because the increase in vegetation coverage would reduce not only the intensity of physical weathering, but also the hydrologic capacity to carry coarse-grains for long distance [10,32]. Since the warming climate promoted OM accumulation and marl precipitation in the lake and prevented upland erosion, the sediments accumulated during the period from 14 to 11 ka consist of two components: (1) a decreasing detrital input, and (2) an increasing OM and carbonate that were produced within the lake. The magnetic susceptibility transition from positive to negative values during the period from 14 to 11 ka suggests that the OM and carbonate

became gradually enriched and dominated the lake sediments by ~ 11 ka.

4.3. The lake level fluctuations in the Holocene (after ~ 11 ka)

The Holocene sediments have very low ferro- and para-magnetic mineral concentration as indicated by the negative magnetic susceptibility values (Fig. 3). Perhaps the most prominent magnetic feature in WL03-1 is the strong ARM and SIRM peaks in the yellowish marl layers, while the intervening gytija layers show the weak ARM and SIRM values (Fig. 3A). This distinct pattern could result from two possible scenarios: (1) comparatively strong dissolution of magnetic minerals in the gytija due to its high OM content while little or no dissolution occurred in the marl layers; or (2) comparatively more detrital input during marl deposition than during gytija deposition.

Dissolution of detrital magnetic minerals has been observed in marine and lake sediments from other regions [10,33] and could have caused the low magnetic mineral concentrations of the gytija layers in WL03-1. Dissolution tends to preferentially remove more fine-grained magnetic particles than coarse-grained magnetic particles because of the higher surface area-to-volume ratios of fine grains [34,35]. At White Lake the very high S-ratios indicate that almost all the magnetic minerals are magnetite with coercivities ≤ 0.3 T. Because the ARMs were applied with a 0.1 T peak alternating field and grain sizes are inversely related to coercivities [36], the ARM-carrying grains with coercivities < 0.1 T are larger in grain size than magnetite grains with coercivities between 0.1 and

0.3 T. The 0.1 to 0.3 T grains would only contribute to the SIRM in our experiments. Step-wise IRM acquisition experiments of selected gytija samples show that the samples are not completely saturated until fields of 0.3 T are reached (Fig. 7A), suggesting that the gytija contains grains with coercivities as high as 0.3 T. Thus, any dissolution of magnetite particles in the gytija would cause a preferential removal of the finest-grained magnetite with the highest coercivities, between 0.1 and 0.3 T, causing a decrease in the SIRM while the ARM is left unaffected. This preferential dissolution would lead to an increase in the ARM/SIRM ratio. To examine whether dissolution has occurred in the gytija, the ARM/SIRM ratios of the marl and gytija were compared. Since the silicate content remained almost constant (~15%) during the Holocene, carbonate content varies inversely with organic matter content. An OM content of 50% was used as a cut-off to consider sediments as gytija for OM > 50% and as marl for OM < 50%. Under this classification scheme, the ARMs and SIRMs of both marl and gytija were analyzed. As shown in Fig. 7B and C, the correlations between the ARMs and SIRMs for both marl

($R^2=0.949$) and gytija ($R^2=0.974$) are excellent. Regression analyses show that the ARM/SIRM ratios are 0.262 for marl and 0.247 for gytija, respectively. A *t*-test was performed to examine whether the ARM/SIRM ratio of gytija is significantly different from that of the marl sediments. The *t*-test yields a $T^*=0.87$, which is less than 1.99 required for $n=66$ ($df=62$) at 95% confidence level, suggesting that the two slopes are not significantly different. Therefore, dissolution probably did not significantly affect the magnetic minerals in the gytija of WL03-1. To examine whether dissolution has occurred in the gytija of WL03-2, the ARM vs. SIRM plots of gytija from both cores were compared. Fig. 7D shows that the ARM/SIRM ratios of the Holocene gytija in WL03-2 are generally similar to those in WL03-1, suggesting that dissolution was not important in WL03-2.

Since dissolution apparently did not significantly affect magnetic minerals in the gytija in WL03-1, the distinct contrast in magnetic mineral concentrations of gytija and marl layers in WL03-1 is probably a result of the increased detrital flux into the lake during the marl deposition, giving rise to the strong ARM and SIRM of the marl layers. It is intriguing that marl yields strong

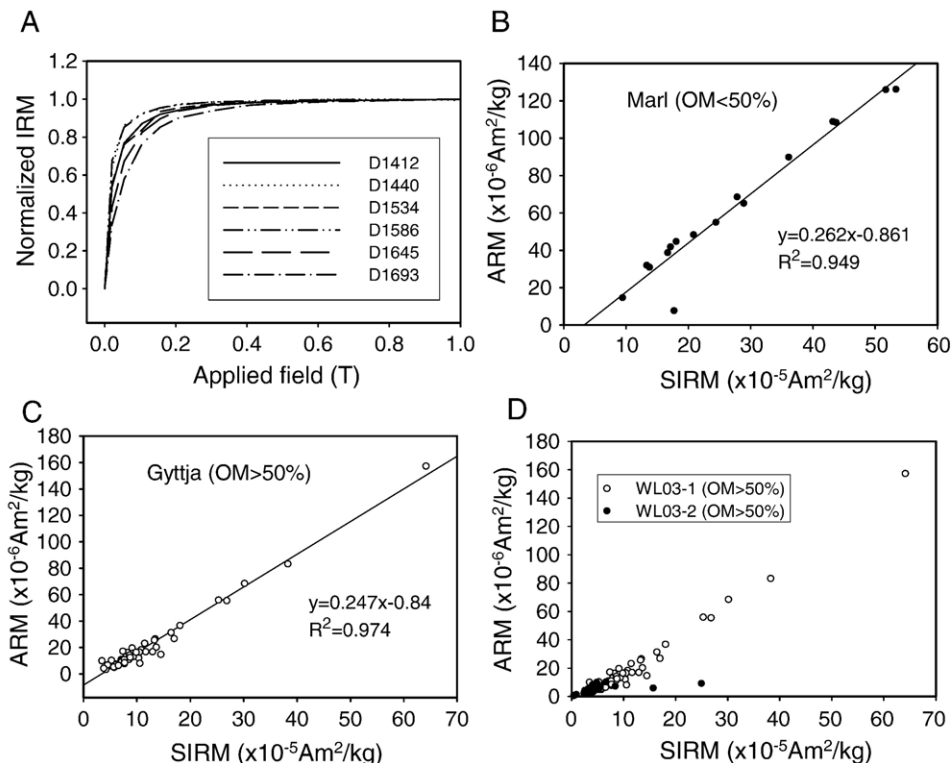


Fig. 7. Step-wise IRM acquisition curves of representative gytija samples (A). The excellent correlations between ARM and SIRM for marls (B) and gytija (C) in core WL03-1 indicate negligible dissolution. The ARM/SIRM ratios of core WL03-2 are generally similar to those of WL03-1 (D), suggesting that dissolution was not important in WL03-2. The two low ARM/SIRM ratio points of core WL03-2 (D) were probably caused by sorting. The open circles shown in (D) are the same data as in (C) for reference.

magnetic signatures because it is believed that marl precipitation in a hardwater lake is mainly induced by photosynthesis of aquatic plants, such as *Chara*, *Potamogeton*, and *Elodea*, or agitation of water by waves. Photosynthesis and agitation of water would remove CO_2 from the lake water, causing the precipitation of CaCO_3 [37,38]. The marl in White Lake is thought to be precipitated largely due to the algae *Chara* [39]. Photosynthesis of *Chara* draws CO_2 into the plants from immediately surrounding water, shifting the bicarbonate–carbonate equilibrium and causing precipitation of carbonate. It is unlikely that the marl precipitation causes magnetic mineral production. Another possibility for the strong magnetization is that the Holocene sediments contain magnetotactic bacteria (MTB), which can produce magnetite. It has been shown that the concentration of MTB-produced magnetite tends to display a positive covariation with OM content [40]. However, our data show no correlations between OM content and magnetic parameters (ARM and SIRM) (Fig. 6C). Also, magnetite grains are probably not derived from MTB based on Moskowitz's [28] criteria (Fig. 6A and D). Therefore, the strong magnetic signatures of marl layers are unlikely to be caused by authigenic production of magnetic minerals. The yellowish marl layers are heterogeneous and contain coarse calcareous particles, indicating that sediments in these marl layers were not produced in-situ [41].

The sediments in these marl layers could be derived from either the upland or the littoral zone. However, it seems unlikely that the increased sediment flux resulted from erosion of the upland soils because the silicate content, a proxy for detritus influx, does not show a corresponding increase during marl layer deposition. Also, the continuous forest vegetation of the region during the Holocene [30] and the lack of permanent inlets at the lake would prevent intense erosion and detritus transportation into the lake. The heterogeneous texture of marl layers could suggest that these layers may represent turbidites. However, this is unlikely for several reasons. The lake is small with a maximum water depth only ~ 14 m (Fig. 1C); Also, the lake is located in a relatively flat terrain and there are no steep slopes around the lake; In addition, the study area is located in a relatively aseismic region [42] and seismic events with strong magnitude are rare [43,44]; Furthermore, the boundaries between marl layers and the underlying gyttja are transitional rather than occurring as abrupt unconformities, thus arguing against a turbidite origin for the marl layers. A more likely explanation is a drop in lake level. Since marl precipitation often occurs in shallow waters where *Chara* are more abundant, low lake levels would allow marl sediments along the shoreline to be exposed and oxidized (Fig. 8). The laboratory exper-

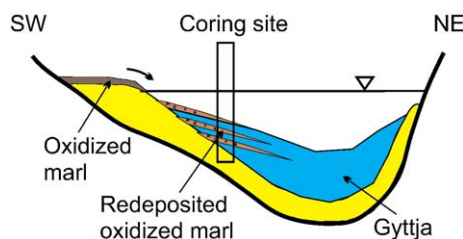


Fig. 8. Schematic diagram showing low lake level periods when precipitated marls near shore are exposed, oxidized, and transported to deeper parts of the lake. SW = southwest; NE = northeast.

iments indicate that oxidation of marl sediments is a feasible way to enhance the magnetization intensities (Table 2). The oxidized marl sediments at the shoreline may be subsequently transported into the lake and deposited as the distinct yellowish marl layers (Fig. 8) that show high magnetic mineral concentrations in WL03-1. Table 2 shows that the post-laboratory oxidation ARM/SIRM ratios are typically lower (0.03–0.08) than those of the marl layers (~ 0.25) in WL03-1, probably due to the limited duration of the experiments that did not allow the magnetic particles to grow big enough to carry an ARM (coercivities < 0.1 T). One possible explanation for the difference in ratios between the naturally oxidized sediments and the laboratory oxidized sediments is that some maghemite was probably produced in the laboratory oxidation experiments. Maghemite has an ARM susceptibility of only about 25% of that of magnetite, but similar coercivities [45], so the laboratory oxidized samples have low ARM/SIRM ratios and almost unaffected S -ratios. The two SIRM peaks (Fig. 3) with low ARM/SIRM ratios at ~ 1.3 and 4.4 ka in WL03-2 (Fig. 7D) can be explained by sorting processes that preferentially carry finer particles with coercivities between 0.1 and 0.3 T further away from source areas during the periods of large lake level drops.

It is not possible to determine the precursor mineral of the secondary magnetic mineral, probably magnetite as indicated by the high- S -ratios (Table 2), in the oxidized marl sediments. Non-magnetic minerals such as siderite and amorphous phases are possible candidates, as oxidation could have converted them into ferromagnetic minerals, leading to the enhanced magnetic mineral concentrations in the oxidized marl sediments.

If the strong magnetism of the Holocene marl layers indicates low lake levels, WL03-1 would record low lake levels at ~ 1.3 , 3.0 , 4.4 , and 6.1 ka, and WL03-2 would record low lake levels at ~ 1.3 and 4.4 ka (Fig. 3). It is intriguing that WL03-1 recorded four low lake levels, while WL03-2 recorded only two low lake levels. The difference in the ability for the sediments at the two

coring sites to record low lake levels may arise from their different locations in the lake. WL03-1 was cored in the relatively shallower part of the lake, where the bottom slopes are gentle and the marl bench is wide (Fig. 1C); whereas WL03-2 was extracted from the deeper part of the lake, where the bottom slopes are steep and the marl bench is narrow (Fig. 1C). Since WL03-1 is closer to a wider marl bench, it is a more sensitive recorder of drops in lake level. Perhaps the lake level drops at ~ 3.0 and 6.1 ka were not large enough to be recorded in the relatively deeper part of the lake where WL03-2 was taken, but large enough to be recorded in the shallower part of the lake where WL03-1 was cored.

Since White Lake has no permanent inlets, the lake level fluctuation of White Lake likely resulted from regional climate change that regulates the moisture availability in the mid-latitude United States. Comparison of the White Lake data with climate records from the North Atlantic sediments shows that low lake levels at ~ 1.3 , 3.0 , 4.4 , and 6.1 ka in White Lake occurred almost concurrently with the cold events at ~ 1.5 , 3.0 , 4.5 , and 6.0 ka in the North Atlantic Ocean [46]. These cold events are associated with the 1500 yr warm/cold cycles in the North Atlantic during the Holocene and the ~ 1500 yr cycle has been interpreted to result from solar forcing [46]. The close correlation between White Lake and the North Atlantic suggests that, in response to the decreased temperatures, the White Lake area climate expressed as periods of reduced moisture abundance [47]. Therefore, the Holocene ~ 1500 yr lake level fluctuations of White Lake probably represent responses to the broad-scale climate variability in the continental North Atlantic region [47].

5. Conclusions and implications

- (1) Two cores from White Lake, New Jersey, show lithologic changes from clays, through marl, to gyttja over the past ~ 18 ka since the ice retreat. Changes in magnetic properties of the clays at ~ 18 – 15 ka provide more detailed information about the deglaciation process and initial development of the lake than lithology data alone would have, for example, in differentiating the lower glacial lacustrine and upper lacustrine clays. From ~ 15 to 14 ka, a gradual increase in organic matter and carbonate indicates the deglacial Bølling warming. This warming trend is accompanied by a decrease in both detrital (silicate) input and magnetic mineral concentration, indicating a productive lake and stabilized watershed

in a continuously ameliorating climate during the latest Pleistocene.

- (2) Since the onset of the Holocene, organic matter content has remained high during the early Holocene. In the middle and late Holocene, sediments alternate between gyttja and marl, particularly in core WL03-1. The marl layers contain coarse particles, display a heterogeneous texture, and have high magnetic mineral concentrations. The elevated magnetic mineral concentration in these marl layers probably resulted from oxidation of precipitated marls near shore during low lake levels and their subsequent erosion, transportation, and redeposition. This interpretation is supported by laboratory heating and open-air exposure experiments. Low lake levels occurred at ~ 1.3 , 3.0 , 4.4 , and 6.1 ka, and probably represent regional responses to well-documented 1500-year Holocene climate cycles.
- (3) This study demonstrates that measurements of multiple magnetic parameters of lacustrine sediments could provide a more detailed and robust interpretation of paleoenvironmental changes than lithology alone would have. Also, laboratory magnetic experiments can provide insight into the underlying processes of lithological changes, which otherwise may not be easy to prove.

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