Late-glacial and Holocene vegetation and climate variability, including major droughts, in the Sky Lakes region of southeastern New York State

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ARTICLE INFO

ABSTRACT

Sediment cores from Lakes Minnewaska and Mohonk in the Shawangunk Mountains of southeastern New York were analyzed for pollen, plant macrofossils, macroscopic charcoal, organic carbon content, carbon isotopic composition, carbon/nitrogen ratio, and lithologic changes to determine the vegetation and landscape history of the greater Catskill Mountain region since deglaciation. Pollen stratigraphy generally matches the New England pollen zones identified by Deevey (1939) and Davis (1969), with boreal genera (*Picea*, *Abies*) present during the late Pleistocene yielding to a mixed *Pinus*, *Quercus* and *Tsuga* forest in the early Holocene. Lake Minnewaska sediments record the Younger Dryas and possibly the 8.2 cal kyr BP climatic events in pollen and sediment chemistry along with an ~1400 cal yr interval of wet conditions (increasing *Tsuga* and declining *Quercus*) centered about 6400 cal yr BP. Both Minnewaska and Mohonk reveal a protracted drought interval in the middle Holocene, ~5700–4100 cal yr BP, during which *Pinus rigida* colonized the watershed, lake levels fell, and frequent fires led to enhanced hillslope erosion. Together, the records show at least three wet–dry cycles throughout the Holocene and both similarities and differences to climate records in New England and central New York. Drought intervals raise concerns for water resources in the New York City metropolitan area and may reflect a combination of enhanced La Niña, negative phase NAO, and positive phase PNA climatic patterns and/or northward shifts of storm tracks.

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1. Introduction

A network of reservoirs and aqueducts originating in New York State's Catskill Mountains supplies New York City (NYC) with ~1.5 billion gallons of drinking water daily (Blake et al., 2000). Despite the dependence of nearly 10 million people on this supply, little is known of the Holocene drought history of the area. Ibe (1985) and Ibe and Pardi (1985) described pollen spectra from two Catskill sites, noting general responses to Laurentide deglaciation and Holocene insolation changes, but no similar studies have been published for the area.

The Catskill region forms the western boundary of New York's Hudson River Valley and lies between well-studied sites in central New York and New England that show contrasting moisture balance since deglaciation (Fig. 1). In central New York, for example, deposition of marls in Cayuga and Owasco Lakes and adjacent fringing wetlands, as well as a variety of proxies in Lake Ontario, indicate that the early

to middle Holocene was a time of warm but wet conditions (Dwyer et al., 1996; Mullins, 1998; McFadden et al., 2005). Cayuga Lake experienced lowstands at 8575, 6830, 4770, and 1955 cal yr BP (7800, 6000, 4200, and 2000 ^{14}C kyr BP), however, all were higher than modern lake level, reflecting overall wetter than modern conditions in the region (Mullins, 1998). In contrast, pollen and lake level evidence from New England (Newby et al., 2000; Shuman et al., 2004) suggests dry climate prior to 8500 cal yr BP, wet conditions from 8500 to 5500 cal yr BP, and widespread drought between 5500 and 3000 cal yr BP, the latter interval likely responsible for the decline of hemlock forests (Miller, 1973; Foster et al., 2006) previously attributed to a pathogen. Post 3000 cal yr BP conditions again became more mesic, with lakes expanding to their highest levels during the Holocene.

Sediment cores and submerged shoreline features from Davis Pond, Massachusetts, reveal even finer-scale variations in lake level in New England superimposed on a general trend toward wetter conditions throughout the Holocene (Newby et al., 2011). Notable lowstands occurred from 13.4 to 10.9 cal kyr BP, at 9.2 and 8.2 cal kyr BP, and from 5.6 to 4.9 cal kyr BP, >3.5 cal kyr BP, 3.0–2.3 cal kyr BP, and 1.6–0.6 cal kyr BP. Several of these lowstands appear to correlate to dry climate cycles recorded in sediment magnetic parameters at White Lake, New Jersey (~6100, 4400, 3000, and 1300 cal yr BP; Li et al., 2007).

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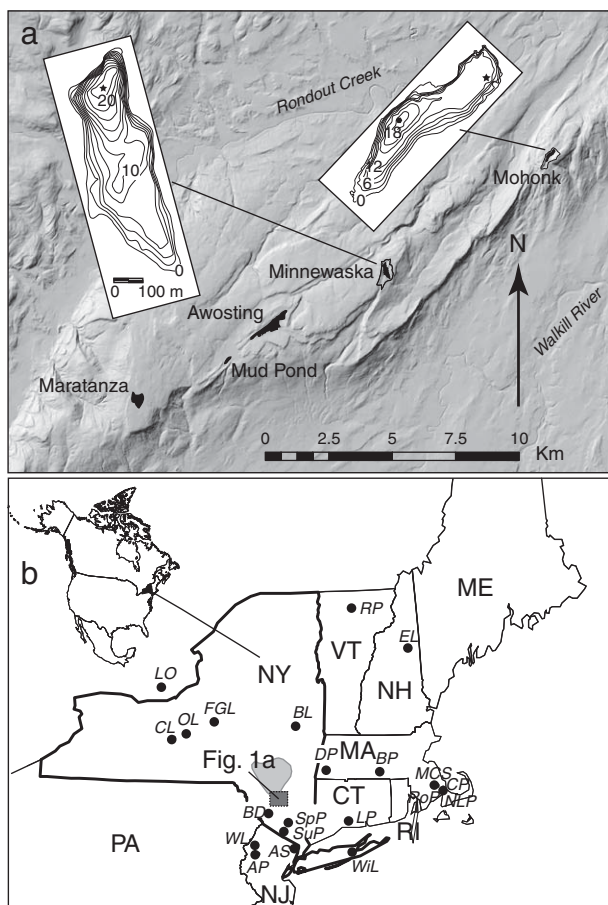Peteet@ldeo.columbia.edu (D.M. Peteet), ryand@unm.edu (R.Y. Anderson).

Fig. 1. Study area. a) Shaded relief map of the Shawangunk Mountains showing the five Sky Lakes, the drainage basins of lakes Minnewaska and Mohonk (black line around each lake), and bathymetric maps of lakes Minnewaska and Mohonk. Depth contours on bathymetric maps are shown in meters. Contour interval for Minnewaska is 2 m and for Mohonk is 3 m. Stars mark the locations where Livingstone cores were taken. Circle denotes an additional Glew core that captured the sediment/water interface. Foothills of the Catskill Mountains are in the northwestern corner of the map. b) Global and regional setting. State labels are as follows: NY=New York, PA=Pennsylvania, NJ=New Jersey, CT=Connecticut, RI=Rhode Island, MA=Massachusetts, VT=Vermont, NH=New Hampshire, and ME=Maine. Black dots show the locations of comparison sites referenced in the text: LO=Lake Ontario, CL=Cayuga Lake, OL=Owasco Lake, FGL=Fayetteville-Green Lake, BL=Ballston Lake, BD=the "black dirt" region, SuP=Sutherland Pond, SpP=Spruce Pond, AS=Alpine Swamp, WL=White Lake, AP=Allamuchy Pond, WiL=Wildwood Lake, LP=Linsley Pond, DP=Davis Pond, BP=Blood Pond, RP=Ritterbush Pond, EL=Echo Lake, MCS=Makepeace Cedar Swamp, CP=Crooked Pond, NLP=New Long Pond, RoP=Rocky Pond. New Long Pond and Rocky Pond are so close to Makepeace Cedar Swamp and Crooked Pond that we have not shown individual symbols for their locations. Location of Fig. 1a is shown in dashed dark gray box. Catskill Mountain region shown in lighter gray shaded region adjacent to the study area, boundary taken from Ibe (1985).

The spatial and temporal variability of climate exhibited in the northeastern United States during the Holocene, along with the importance of the Catskill region for New York City's water supply, argues for additional documentation of the paleoecology and paleoclimate of the area. Here we report on a multi-proxy climate record that spans the period of deglaciation through the late Holocene obtained from sediment cores in two lakes (Minnewaska and Mohonk) in the Shawangunk Mountains of southeastern New York, a range immediately adjacent to the Catskill Mountains with an exceptionally detailed and long phenological record correlated to climate (Cook et al., 2008). Questions addressed in our study include whether the early to middle Holocene interval showed wetter or drier than modern conditions, whether drought intervals resemble those in New England or central New York, and whether a climatic driver led to mid-Holocene *Tsuga* decline as proposed for western New York (Miller, 1973) and New England (Foster et al., 2006).

2. Study area

Lakes Minnewaska and Mohonk lie atop the Shawangunk Mountains in southeastern New York, ridges of Silurian-age quartz pebble conglomerate and quartz sandstone (Shawangunk Formation, 99% quartz) that overlie Ordovician-age calcareous shales (Martinsburg Formation) (Fig. 1; Macchiaroli, 1995; Bernet et al., 2007). Bedrock occurs either at the surface or within 30 cm of the surface over more than 70% of the range, reflecting glacial stripping and extremely slow rates of soil formation (Coates et al., 1994; Laing, 1994). During the late Pleistocene, ice from the Hudson-Champlain lobe of the Laurentide ice sheet covered the mountains to a depth >600 m (Coates et al., 1994). Glacial plucking of tectonically fractured bedrock produced five lake basins trending in a NE–SW direction along the ridge crest, collectively known as the "Sky Lakes" (Fig. 1). Steep bedrock walls and small drainage areas surround each lake. Water chemistry reflects the depth of glacial plucking: Lakes Awosting, Maratanza, and Mud Pond lie entirely within the Shawangunk formation and are acidic, whereas Mohonk Lake bottoms in the Martinsburg Formation and has neutral pH (Schiff, 1986; Friedman et al., 1989). Lake Minnewaska lies near a small structural inlier of the shale in the conglomerate, and shows a pH higher than the three southern lakes but more acidic than Mohonk Lake.

Reflecting their ridge top positions, neither Mohonk (380 m elevation, 7.4 ha, 27% of catchment area) nor Minnewaska (500 m elevation, 11.5 ha, 28% of catchment area) has a perennial inflow stream, thus both are fed primarily by direct precipitation supplemented by groundwater flow introduced through fractures and by sheet wash from surrounding rock walls (Schiff, 1986; Coates et al., 1994). Two very short (30–50 m long) and low flow (<10% of total inflow) ephemeral streams originating in springs around the southern margin of Mohonk Lake provide runoff during the spring snowmelt season (Herczeg, 1985; Schiff, 1986). Lake Minnewaska overflows at its southern end through a natural outlet and also leaks water to bedrock fractures along its southern and northwestern margins (Friedman et al., 1989). Mohonk Lake presently drains to the north through an outlet controlled by the Mohonk Mountain House resort, but would also have drained northward through bedrock fractures prior to development (Herczeg, 1985; Schiff, 1986; J. Thompson, personal communication). Water residence time in Mohonk Lake was determined to be approximately 4.5 years (Herczeg and Fairbanks, 1987). No similar study has been done for Minnewaska, but given its volume (1.064∗10^9 L; Coates et al., 1994) and the rate of flow in its outlet stream (10–37 L/s in summer baseflow measured ~3.5 km downstream of the lake; C. O'Reilly, personal communication), residence time is unlikely to be greater than a decade or two.

The lack of inflow streams means that sedimentation rates are low and that deposition is dominated by organic materials falling or blowing into the lakes from the surrounding cliffs or produced within the lakes by primary production. The drainage area of Lake Minnewaska presently contains chestnut oak forest with lesser amounts of hemlock-northern hardwood forest (Thompson, 1996) growing upon very thin soils developed in low permeability glacial till or high permeability glaciofluvial sediments (Coates et al., 1994). The vegetation surrounding Mohonk Lake consists primarily of hemlock-northern hardwood forest, with small amounts of chestnut oak and other deciduous forest (Thompson, 1996). These botanical communities, along with adjacent pitch pine barrens, Appalachian oak-hickory forest, and beech-maple mesic forest support the highest biological diversity in New York State and contribute to the Shawangunk ridge being designated by the Nature Conservancy as one of the "Last Great Places" on earth (Thompson and Huth, 2011). In the lakes, aquatic producers include cyanobacteria, diatoms, chrysophaceae, and chlorophyceae (Hellerman, 1965; Herczeg, 1985).

Staff at Mohonk Preserve, a private non-profit land conservancy surrounding Mohonk Lake, operate the Mohonk Lake Cooperative Weather Station, which is part of the National Oceanographic and Atmospheric Administration network. A remarkable and unique set of weather records have been collected daily for 114 years by a limited

number of observers and with no change in instrumentation or methodology (Cook et al., 2010). Average January and July temperatures over the entire record measure −4 °C and 21.5 °C respectively (J. Thompson, personal communication), though the record also shows the warming trend that occurred throughout the 20th century and continued into the 21st (0.013 °C/yr; Cook et al., 2010). Additional records, such as timing of first and last ice on Mohonk Lake, number of freeze days, and various phenological records indicate that the area is sensitive to regional climatic trends and does not merely reflect local conditions (Cook et al., 2008, 2010). Precipitation is fairly uniform throughout the year, ranging from a minimum of 83.5 mm in February to a maximum of 114.8 mm in July, and having an annual mean value of 1215.4 mm. Potential evapotranspiration measures 650 mm (Coates et al., 1994).

The very thin soils (sometimes only a few cm thick; Laing, 1994) on the Shawangunk ridge and the fact that they are either very poorly (glacial till origin) or excessively (glaciofluvial sands) drained means there is little soil moisture storage capacity (Coates et al., 1994). Combined with the low porosity and permeability of the underlying quartz pebble conglomerate, the soil characteristics ensure high rates of runoff and make the area vulnerable to drought.

3. Methods

3.1. Core extraction

Sediment cores were extracted from Lakes Minnewaska and Mohonk in winter using a modified Livingstone piston corer (Wright et al., 1984) dropped through a hole in the frozen lake surface (Fig. 1). Extruded cores were wrapped in plastic wrap, placed in PVC pipe, and refrigerated prior to sampling. Cores were split and described the day after extraction, followed by sampling. An additional gravity core was taken from Mohonk Lake in summer using a Glew corer (Glew, 1991) dropped from a boat. While not analyzed for paleoclimatic reconstruction here, this core yielded additional chronological information.

3.2. Laboratory analyses

3.2.1. Dating

Lake Minnewaska samples for accelerator mass spectrometer (AMS) radiocarbon dating consisted of bulk organic sediments whereas Mohonk Lake samples consisted of either plant macrofossils or bulk sediments. In the case of bulk sediments, samples were acidified to remove any potential carbonate, rinsed, dried, and ground. Samples were analyzed by Beta-Analytic Laboratories. The IntCal09 calibration curve was used to convert from radiocarbon to calendar ages (Reimer et al., 2009).

3.2.2. C/N ratio, $\delta^{13}C$, and organic carbon content

One cubic centimeter samples for organic carbon content and stable isotopic analysis were taken every 3–5 cm along each core. Samples were treated with dilute HCl to remove any carbonate, rinsed in deionized water, dried, and powdered. Organic carbon contents are proxies for terrestrial vegetation inputs and lacustrine primary production (Ellis et al., 2004) and were measured with a UIC Inc., 5012 Coulometer. C/N ratios and carbon isotopic values are used to distinguish terrestrial versus lacustrine organic matter (Meyers, 1994; McFadden et al., 2005) and were determined using a Carlo Erba NA 1500 Series II NC elemental analyzer connected to a GV Instruments Optima stable isotope mass spectrometer at the University at Albany in Albany, NY.

3.2.3. Pollen

Additional 0.5-cm³ samples were taken for pollen analysis and washed through a 250 μm sieve to separate out large plant fragments. The sieved samples were then subjected to 10% HCl to remove carbonates, concentrated HF to remove silicates, and 10% KOH to remove organic matter (Moore et al., 1991). After processing, pollen grains were mounted in glycerin and identified using keys (Erdtman, 1952–1965; Kapp et al., 2000) and a modern reference collection at Lamont Doherty Earth Observatory. Counting was done at 400× magnification, and a minimum of 300 grains were counted per sample. Unknowns constituted less than 4% of any particular sample. Vegetation zone boundaries were identified by visual inspection of both pollen and macrofossil data (see below) and confirmed using stratigraphically constrained cluster analysis of pollen data by the method of incremental sum of squares (CONISS; Grimm, 1987).

3.2.4. Macrofossils and charcoal

Samples ~3 cm³ in volume were used for macrofossil and macroscopic charcoal analysis following the procedure of Watts and Winter (1966). Macrofossils are assumed to indicate the local presence of vegetation, as opposed to pollen, which may be either locally or regionally sourced. Charcoal is used to reconstruct fire history. Each sample was washed through a 500 μm sieve, and the materials retained on the sieve identified and counted in water at about 60× magnification. Samples were identified using an extensive Lamont Doherty Earth Observatory reference collection and keys (Martin and Barkley, 1973; Montgomery, 1977; Crum and Anderson, 1981).

4. Results

4.1. Lake Minnewaska

4.1.1. Core description

The Lake Minnewaska core (Fig. 2) measures 2.4 m in length and was taken in a depth of >20 m of water (Fig. 1). The bottom ~0.6 m of the core (250–190 cm) consists of layered silts and sands, with individual layers a few mm in thickness. Above this unit and in abrupt transition with it lies ~0.5 m of massive light gray clay (190–140 cm depth), and both units are nearly devoid of organic matter (Fig. 3). A gradual transition to organic-rich sediments begins between 140 and 135 cm, and by 120 cm, organic carbon content has climbed from <1% to >20% of sediment mass, coincident with a 7‰ shift to more negative values of $\delta^{13}C$. As organic content rises, the core gradually transitions from a medium gray clay between 135 and 125 cm, to a clayey gyttja from 125 to 85 cm, to black gyttja from 85 cm to the top. Within the three organic-rich horizons, large oscillations in organic carbon content and carbon to nitrogen ratio (C/N) are superimposed on long-term trends toward increasing values. $\delta^{13}C$ values vary by ~2‰ across the same horizons, measuring between −30 and −28‰.

4.1.2. Pollen and macrofossil stratigraphy

Pollen concentrations below 140 cm depth are too low to allow statistically significant counts, so we truncate the pollen diagram at this depth (Fig. 2). Shrub, herb, and aquatic pollen were counted but were very low in abundance and did not help illuminate climatic shifts. Accordingly, we focus on the arboreal taxa, showing those genera that constitute 5% or more of the total pollen sum in any particular sample.

Zone Min-1 begins at 140 cm, coincident with the steep rise in organic carbon content (Fig. 2), and extends to 125 cm. *Pinus* (pine) is dominant, constituting on average 40% of the total pollen sum, with values as high as 50%. *Picea* (spruce), *Abies* (fir), *Quercus* (oak), and *Betula* (birch) are also prominent (10–20% of the total depending on genus). *Ostrya-Carpinus* (hornbeam) contributes about 8% of the total sum. *Fraxinus* (ash) pollen rises from ~0.5 to 3.5% across the zone. Plant macrofossils reflect the presence of conifers and *Betula glandulosa* (shrub birch) in the watershed and *Potamogeton* (pondweed) in the lake (Fig. 4).

Zone Min-2 (125–112 cm) shows a decline in *Pinus* pollen abundance to 30% and declines in *Picea*, *Abies*, and *Quercus* to 10%, 7%, and 7% respectively (Fig. 2). *Alnus* (alder) pollen percentage rises from 0% in Zone Min-1 to 8% here. *Betula* also expands dramatically, reaching 30% of the total pollen sum by the middle of the zone. *Ostrya-Carpinus* declines to about 5% of the total and *Tsuga* (hemlock) rises to about 5% by the end of the zone.
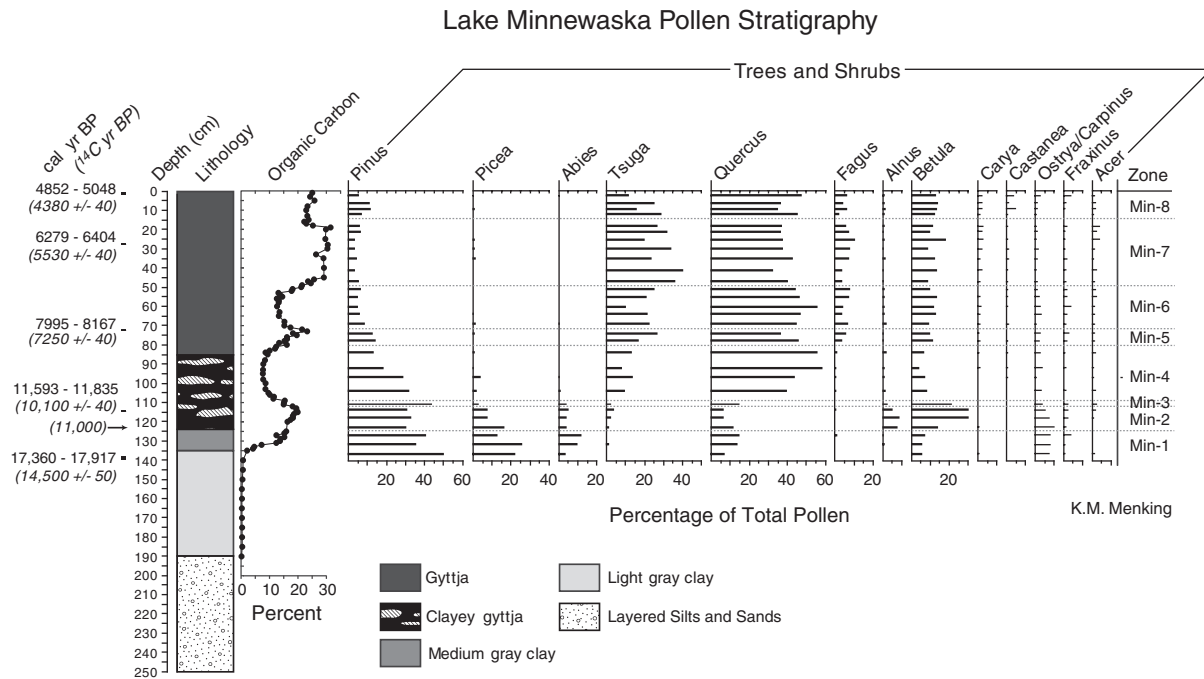
Lake Minnewaska Pollen Stratigraphy

Fig. 2. Lithology and pollen stratigraphy of the Lake Minnewaska core. Sediments below 140 cm are nearly devoid of organic matter and yielded insufficient pollen grains for analysis. Depths of AMS radiocarbon dates and organic carbon content are also shown. Bold italicized date in parentheses and denoted with an arrow marks the regional onset of the Younger Dryas cooling event.

Acer (maple) appears in the middle of the zone and constitutes ~1% of the total sum. *Fraxinus* declines to about 2%. Plant macrofossils indicate the presence of *Picea, Tsuga,* and *Betula papyrifera* (paper birch) in the watershed (Fig. 4). A reddish aquatic bryophyte (moss), *Fontinalis,* with elongated internodes, occasionally present earlier in the core, expands and is found as a thick mass in the core.

Zone Min-3 (112–110 cm) is marked by a very brief (one sample) return to high *Pinus* concentrations (45%), with concomitant declines in

Picea, Abies, and *Tsuga* pollen abundance (Fig. 2). *Alnus* and *Betula* also decline. *Quercus* pollen rises to ~15%. *Tsuga* branchlets and abundant conifer leaf scales (probably *Pinus strobus* — eastern white pine) are found in this zone, along with macroscopic charcoal (Fig. 4). The bryophyte remains present but declines in abundance.

Zone Min-4 (110–80 cm) continues the *Quercus* pollen rise, reaching ~55% by the end of the zone (Fig. 2). At the same time, *Tsuga* pollen abundance undergoes an expansion to 15% while *Pinus* declines to the same
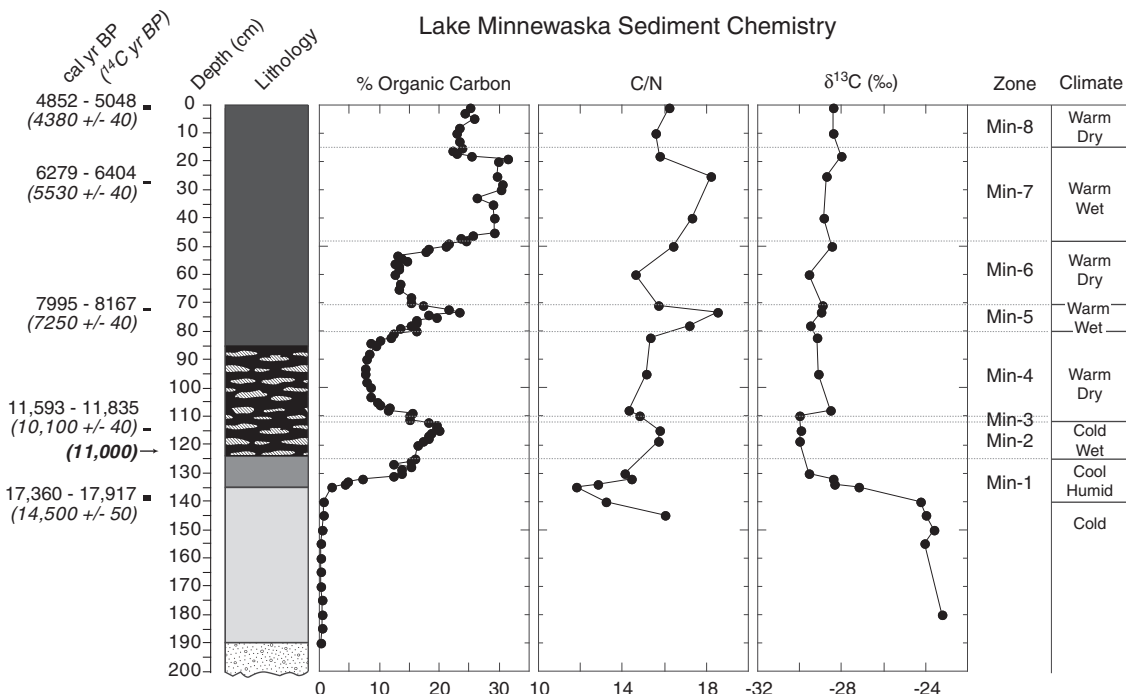

Fig. 3. Chemistry of Lake Minnewaska sediments. C/N refers to carbon to nitrogen ratio. No analyses were performed below 200 cm depth due to low organic carbon contents, and N values below 145 cm depth were too low to allow C/N determination.
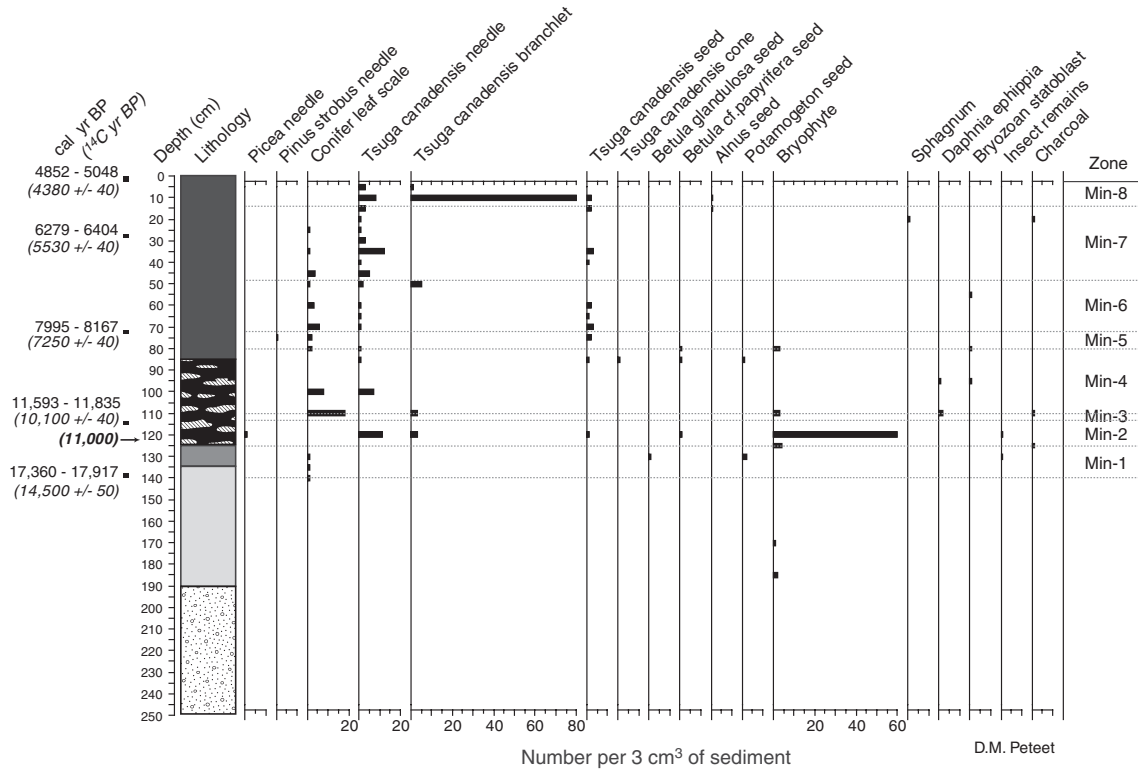
Lake Minnewaska Plant Macrofossils

Fig. 4. Macrofossils and charcoal found in the Lake Minnewaska core.

value. *Picea* declines to 1% and *Abies* disappears. *Betula* and *Alnus* decline to ~7% and ~1% of the total sum respectively. *Ostrya-Carpinus* declines to about 3%. *Castanea* (chestnut) and *Carya* (hickory) make their first appearances in the middle of this zone, measuring ~1%. *Fraxinus* pollen declines to 1% while *Acer* rises slightly (~1.5%). *Tsuga* and *Betula papyrifera* macrofossils appear along with *Potamogeton*, *Daphnia* (water flea), and bryozoans (moss animals) (Fig. 4).

Zone Min-5 (80–72 cm) continues the *Tsuga* pollen expansion and *Pinus* pollen decline, reaching ~27% and ~12% respectively by the end of the zone (Fig. 2). *Quercus* declines to ~36%, and *Castanea* and *Alnus* disappear. *Fagus* increases from <1% of the total sum in zone Min-4 to >4% here. *Fraxinus* pollen abundance expands slightly. *Tsuga* seeds and conifer leaf scales are found in this zone, with a *Pinus strobus* needle found near 73 cm (Fig. 4).

Zone Min-6 (72–48 cm) shows a rebound in *Quercus* to higher values, reaching 55% of the total sum midway up the zone, whereas *Tsuga* declines to 15% at this same depth before rebounding toward the end of the zone. *Fagus* (beech) and *Betula* pollen percentages oscillate between 5 and 10% and 5 and 15% of the total sum respectively. *Carya*, *Ostrya-Carpinus*, and *Acer* show percentages similar to those in zones Min-4 and Min-5, while *Castanea* and *Alnus* return at low abundance (<1% of the total sum). *Tsuga* needles and seeds and conifer leaf scales are common throughout this zone (Fig. 4).

Zone Min-7 (48–15 cm) shows maximal expansion in *Tsuga* pollen to ~40% at 40 cm, though values generally oscillate between 20 and 35% (Fig. 2). *Quercus* declines from its maximum in zone Min-6, averaging about 40% of the total sum. *Pinus* measures minimal values for the Holocene at 5%. *Fagus* and *Betula* reach maximal values for the Holocene in this zone while *Carya* expands slightly to ~2.5%. *Ostrya-Carpinus* and *Fraxinus* pollen percentages both decline slightly relative to their abundances in zone Min-6, while *Acer* increases. *Tsuga* needles are found throughout the zone and in greater abundance than in Min-6 (Fig. 4). *Tsuga* seeds and conifer leaf scales are also found. Macroscopic charcoal appears near the top of the zone along with *Sphagnum* (moss).

Zone Min-8 (15–0 cm) begins with an expansion of *Castanea* pollen abundance from ~0.5% to 4% (Fig. 2). *Pinus* pollen, which had fallen to 5% in zone Min-7, expands to >10% in Min-8, only to decline to 5% again at the top of the core. *Tsuga* pollen declines from a maximum of 25% mid-zone to 12% at the top of the core. *Quercus* pollen expands from 35 to 48% of the total sum. Large numbers of *Tsuga* branchlets are found in this zone along with *Tsuga* needles and seeds (Fig. 4).

4.1.3. Radiocarbon dates

Five AMS radiocarbon dates constrain the timing of major changes in sediment chemistry and vegetation (Table 1; Figs. 2–4). All dates are in chronological order and show a trend toward compaction with depth (Fig. 5). Maxima in organic carbon content and C/N ratio occur at 11,593–11,835, 7995–8167, and 6279–6404 cal yr BP.

4.2. Mohonk Lake

4.2.1. Core description

The Mohonk Lake core (Fig. 6) measures 2.1 m in length and was taken in a water depth of 13.7 m. The core bottoms in sand (211–208 cm depth), above which lies ~0.6 m of gray organic-poor clay (208–144 cm). As in the case of the Minnewaska core, the transition from sand to clay is abrupt. A 0.12-m-thick mat of red stems (aquatic bryophyte *Fontinalis*) caps the clay at 144 cm depth. Above this mat lie silty clay (132–119 cm), poorly sorted sand (119–112 cm), silt (112–108 cm), clayey gyttja (108–102 cm), poorly sorted sand (102–87.5 cm), sandy gyttja (87.5–78 cm), clayey gyttja (78–50 cm), silty clay (50–42 cm), gyttja (42–11 cm), and silty clay (11–0 cm). Transitions between adjacent units are either abrupt or gradational. For example, the transition from clayey gyttja to poorly sorted sand at 102 cm is abrupt, whereas the transition from poorly sorted sand to sandy gyttja and then to clayey gyttja that occurs from 102 to 50 cm is gradual. Organic carbon content in the core varies from nearly 0% in the basal sand and clay units to a maximum of 20% in the bryophyte mat

Table 1
Radiocarbon dates for the Minnewaska and Mohonk cores.

Lab #	Depth (cm)	Sample type	Conventional ^{14}C yr BP ± SD	Calibrated calendar age at 2σ[a]	δ^{13}C (‰)
Minnewaska					
Beta-148365	0–2	1 cm³ sediment	4380 ± 40	4852–5048 (98.7%)	−28.9
Beta-148366	27–28	1 cm³ sediment	5530 ± 40	6279–6404 (100%)	−27.8
Beta-148367	72–73	1 cm³ sediment	7250 ± 40	7995–8167 (97.9%)	−28.8
Beta-148368	114–115	1 cm³ sediment	10,100 ± 40	11,593–11,835 (71.6%)	−29.5
Beta-148369	138–140	1 cm³ sediment	14,500 ± 50	17,360–17,917 (95.4%)	−24.0
Mohonk					
Beta-246219	3–4	Conifer needle	1210 ± 40	1055–1263 (98.5%)	−26.7
Beta-247997	72–73	1 cm³ sediment	3330 ± 40	3445–3635 (100%)	−26.9
Beta-246220	125–126	1 cm³ sediment	7700 ± 40	8410–8557 (99.1%)	−28.2
Beta-246221	139	Plant fragments	11,590 ± 40	13,300–13,591 (100%)	−34.9
Adjacent Mohonk Core					
Beta-181978	5–6	1 cm³ sediment	110.3 ± 0.4 pMC	Modern	−30.0

pMC refers to percent modern carbon.
 [a] Calendar ages determined using IntCal09 (Reimer et al., 2009). Probability that the calendar age falls within the 2σ range given by percentages in parentheses.

(Fig. 7). Above the mat, organic carbon declines to values between 0 and 16%, with sand units showing low values and gyttja-rich layers revealing higher values. Unlike in the Minnewaska core, carbon isotopes on organic matter show little variation with depth with the exception of the bryophyte mat, which is nearly 10‰ more negative than the rest of the core. C/N ratios broadly mirror variations in organic carbon content.

4.2.2. Pollen and macrofossil stratigraphy

The Mohonk core is nearly devoid of pollen below 1.5 m depth, so we have truncated the pollen diagram accordingly (Fig. 6). Zone Moh-1 is represented by one sample below 148 cm in which the pollen spectrum consists solely of *Pinus* and *Picea*. This zone is devoid of plant macrofossils but contains some insect remains and pebble and sand sized clastic sediment grains (Fig. 8).

Zone Moh-2 (148–133 cm) contains the first appearance of *Quercus* pollen in the pollen stratigraphy (Fig. 6). Values for *Abies, Alnus, Betula,*

and *Ostrya-Carpinus* pollen range between 1 and 10% of the total sum. *Pinus* and *Picea* pollen values are high, together making up ~60% of the assemblage. The reddish aquatic bryophyte *Fontinalis* forms a dense mat in this zone, and *Daphnia* ephippia (egg cases) and *Picea* macrofossils are present (Fig. 8). Rocks and sand are absent except in the basal sample.

Zone Moh-3 (133–128 cm), represented by only one sample at 131 cm, shows a major rise in *Alnus* pollen abundance from 1 to 6% (Fig. 6). *Betula* pollen more than doubles to become 7% of the total sum, *Abies* increases to nearly 4%, and *Picea* pollen also expands. *Pinus, Quercus,* and *Ostrya-Carpinus* values decline slightly. Plant macrofossils are the same as those found in Moh-2, including *Picea* and the thick aquatic bryophyte mat (Fig. 8). The lack of rocks and sand is notable.

Zone Moh-4 (128–112 cm) is characterized by a large rise in *Quercus* pollen to ~40% (Fig. 6). *Tsuga* pollen abundance increases to ~15% from a value of <1% in Moh-3, and *Betula* rises to ~7%. *Pinus* declines to ~20%, and *Picea* and *Abies* fall to ~1%. *Carya* and *Acer* pollen first appear in this zone. *Fagus* pollen abundance expands from its first appearance in Moh-3 to make up ~3% of the sum. Conifer leaf scales are present, and charcoal first appears (Fig. 8). Pebbles and sand are also present in the upper portion of the zone.

Zone Moh-5 (112–102 cm) continues the *Quercus* pollen expansion to 50% and the *Pinus* decline to less than 20% (Fig. 6). *Fagus* pollen abundance expands slightly to ~4% while *Betula, Carya,* and *Ostrya-Carpinus* remain about the same as in Moh-4. *Acer* constitutes ~1.5% of the total sum. Plant macrofossils show that conifers and *Tsuga* are present in the overstory surrounding the lake (Fig. 8).

Zone Moh-6 (102–78 cm) is marked by the sustained appearance of *Pinus rigida* (pitch pine) needles, which remain present until the base of Moh-7 (Fig. 8). Additional macrofossils include *Tsuga* seeds and cones, the seeds of *Vitis* (grape), *Rubus* (bramble bushes such as blackberry), *Compositae,* and the seeds and cone scales of *Betula.* In the lake, *Najas* seeds become more abundant and are joined by *Chara* and *Nitella* (both green algae stoneworts). Bryozoan statoblasts, charcoal, and sand are also common in the >500 μm size fraction. *Tsuga* pollen shows a five-fold decline in percentage abundance mid-zone, from ~15 to ~3% of the total sum (Fig. 6), coincident with the highest abundances of *Najas, Chara,* and *Bryozoan* macrofossils and the highest abundance of macroscopic charcoal. *Acer* shows a similar decline from ~4 to ~1% across the zone. While *Tsuga* and *Acer* pollen decline, *Castanea* pollen increases sharply from <1% mid-zone to ~25% by the end of the zone. *Pinus* pollen shows a slight increase across the zone.

Zone Moh-7 (78–38 cm) is characterized by a recovery of *Tsuga* pollen abundance to ~7%, a decline in *Castanea* to about 5%, and a decline in *Quercus* to ~31% of the total sum (Fig. 6). *Fagus* expands slightly from 3% in zone Moh-6 to ~5% here. *Pinus, Alnus, Fraxinus,* and *Acer* show similar values to zone Moh-6, and *Betula* is fairly constant at ~9%.
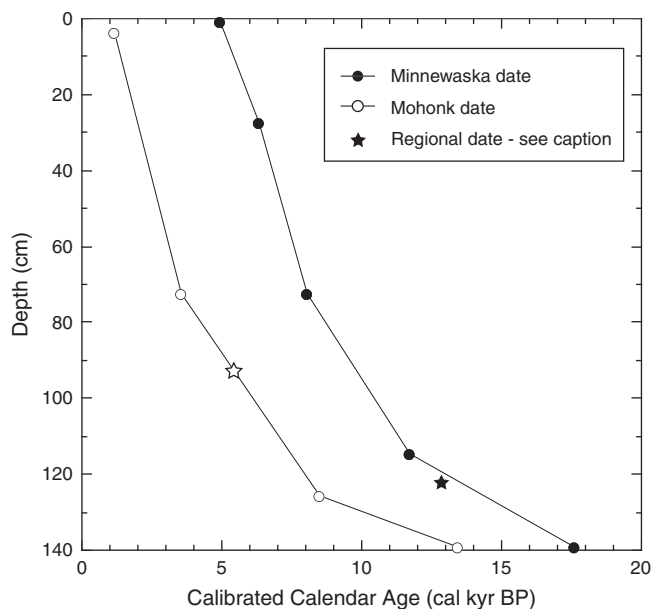

Fig. 5. Age models for the Minnewaska and Mohonk cores. Both cores display compaction with depth. Stars show locations of dates acquired by correlation to the regional pollen stratigraphy. Solid star on the Minnewaska curve is the calendar age of the onset of the Younger Dryas cooling event. Open star on the Mohonk curve is the calendar age of the onset of *Tsuga* decline.
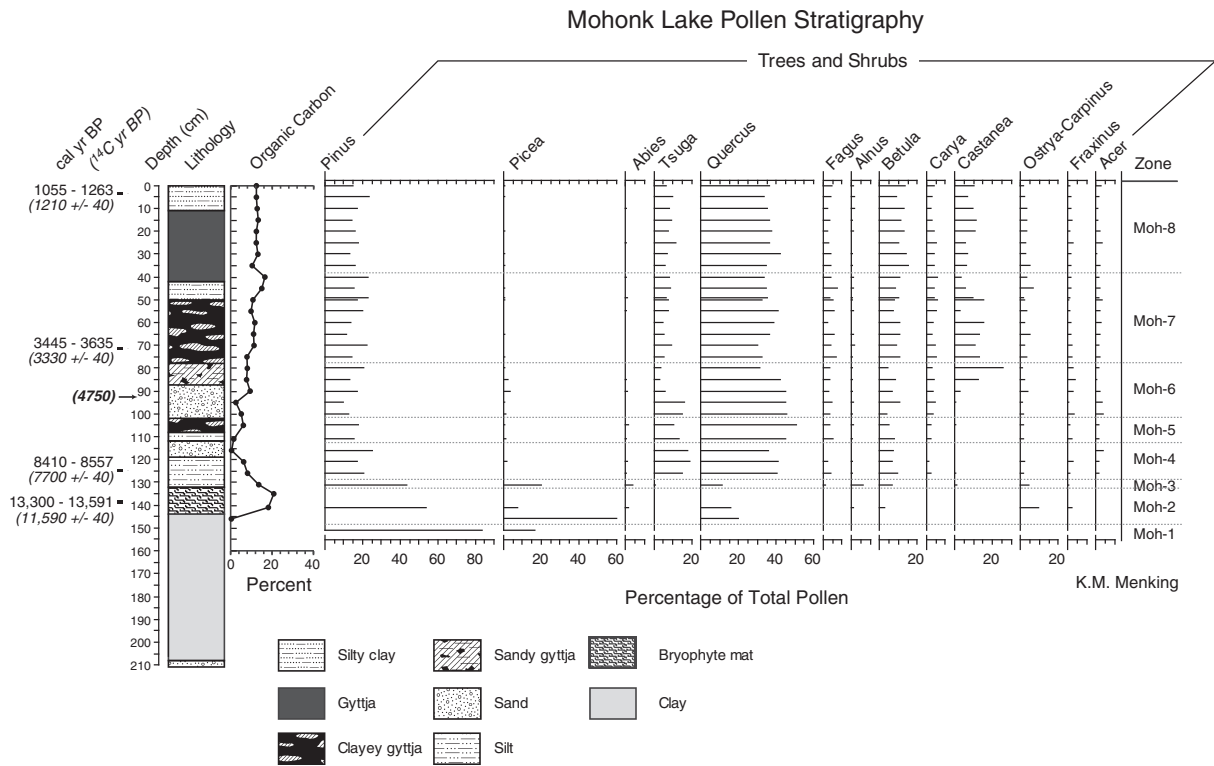
Mohonk Lake Pollen Stratigraphy

Fig. 6. Lithology and pollen stratigraphy of the Mohonk Lake core. Sediments below 150 cm are nearly devoid of organic matter and yielded insufficient pollen grains for analysis. Depths of AMS radiocarbon dates and organic carbon content are also shown. Bold italicized date in parentheses and denoted with an arrow marks the regional onset of *Tsuga* decline.

Macrofossils indicate that *Pinus rigida* disappears at the base of the zone, while *Pinus strobus* is present near the top. *Najas, Chara, Nitella*, and bryozoans also disappear in this zone, which is marked by much lower charcoal abundances than zone Moh-6 and a complete absence of rocks and sand (Fig. 8).

Zone Moh-8 (38–0 cm) contains a rise in *Betula* pollen abundance to ~11% (Fig. 6). *Quercus* pollen holds steady at around 37%, and *Pinus* averages ~17%. *Tsuga, Alnus, Ostrya-Carpinus, Carya, Fraxinus*, and *Acer* show values similar to those in zone Moh-7. *Castanea*, which had been declining in zone Moh-7, rebounds in zone Moh-8 to ~10% of the total sum. *Fagus* pollen abundance declines slightly to ~3.5%. *Betula* seeds and cone

scales along with conifer leaf scales and *Pinus strobus* needles are found, but no macrofossils of *Tsuga* appear. A few isolated macrofossils of *Carex* (sedge), *Najas*, and *Ranunculus aquatilis* (white-water buttercup) occur. Bryozoan statoblasts return, and charcoal increases somewhat. In the >500 μm size fraction, rocks and sand re-appear.

4.2.3. Radiocarbon dates

Four AMS radiocarbon dates constrain the timing of major changes in sediment chemistry and vegetation at Mohonk Lake (Table 1; Figs. 6–8). As in the Minnewaska core, all dates are in chronological order and show a trend toward compaction with depth (Fig. 5). A
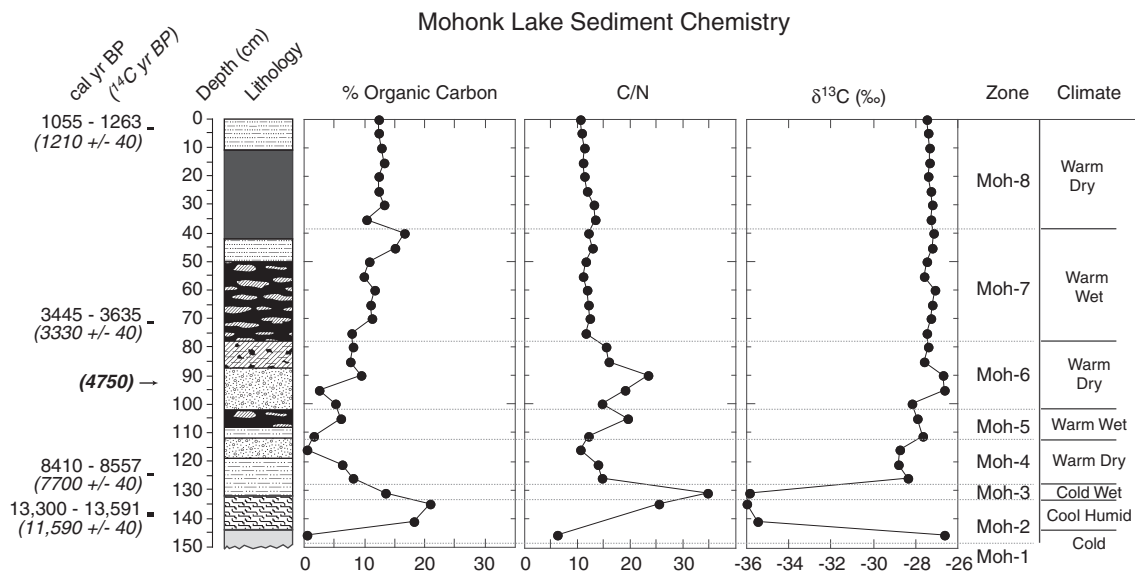
Mohonk Lake Sediment Chemistry

Fig. 7. Chemistry of Mohonk Lake sediments. No analyses were performed below 148 cm depth due to low organic carbon contents. C/N refers to carbon to nitrogen ratio.
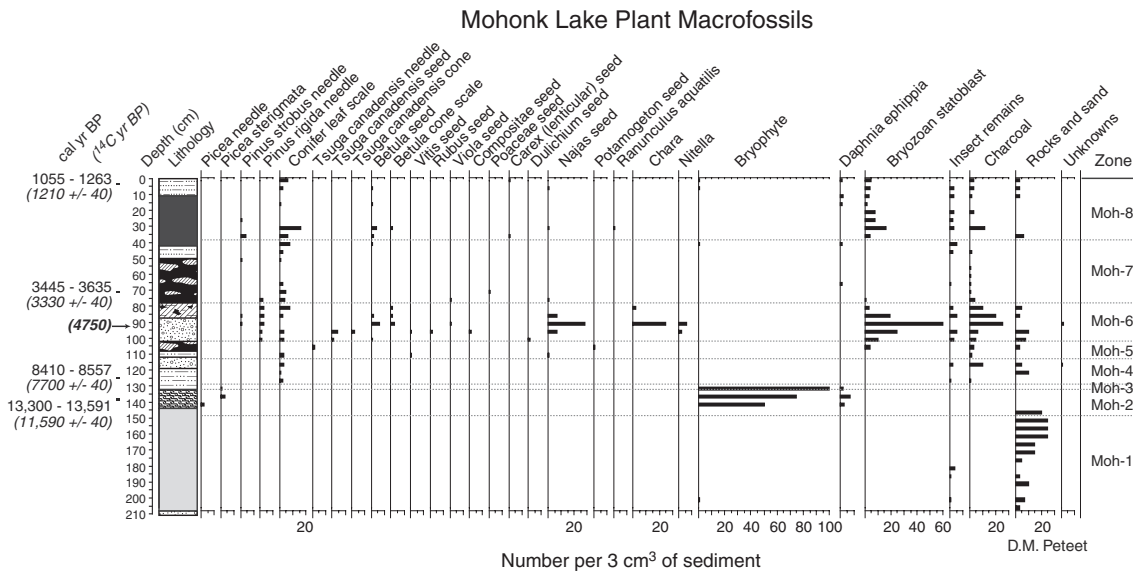
Mohonk Lake Plant Macrofossils

Fig. 8. Macrofossils and charcoal found in the Mohonk Lake core.

sample taken from the middle of the bryophyte mat dates to 13,300–13,591 cal yr BP (11,590 ^{14}C yr BP).

5. Discussion

5.1. Age control

The dates at the tops of the Minnewaska and Mohonk Livingstone cores (Table 1, Fig. 5) indicate that both cores over-penetrated the sediment/water interface. Alternatively, the older ages at the top of each core could be related to reservoir effects or to contamination of younger sediments with older materials.

Three lines of evidence argue against a reservoir effect for either lake. First, the Glew core taken from Mohonk Lake (circle in Fig. 1) captured the sediment/water interface and yielded a modern radiocarbon age at the interface (Table 1). If the calcareous shales of the Martinsburg Formation do not impact the radiocarbon ages of aquatic producers contributing to Mohonk Lake sediments, it is unlikely that bedrock dissolution affects Lake Minnewaska, since it lies entirely within quartz pebble conglomerate, with just a small inlier of shale in its watershed. Second, the residence time of water in each lake is too short to cause isotopic disequilibrium between CO_2 in the atmosphere and that dissolved in lake water. Third, bulk organic sediment dates are consistent with the timing of regional pollen and macrofossil events, as will be demonstrated below.

The first and third lines of evidence also argue against contamination of younger sediments with older materials. An exception to this argument is the basal date on the Minnewaska core, which gives the onset of organic deposition at 17,360–17,917 cal yr BP. This date appears to be in conflict with regionally younger ages for deglaciation and forest arrival based on plant macrofossils (Peteet et al., 2012) and may reflect mixing of older organics contained within clastic sediments derived from glacial tills with younger organics produced by lacustrine primary producers and leaf litter.

The chronology presented below is based on a composite of the Minnewaska and Mohonk cores supplemented by two dates derived from correlation with the regional pollen stratigraphy. These dates are the time of the onset of the Younger Dryas (revealed by the rise in *Alnus* and *Betula* in zone Min-2), which we take to be ~12,920 cal BP based upon nearby records (Peteet et al., 1990, 1993; Maenza-Gmelch, 1997b; Toney et al., 2003; Ellis et al., 2004; Shuman et al., 2004) and the onset of the mid-Holocene hemlock decline (revealed by a decline in *Tsuga* pollen in zone Moh-6), currently dated at 5520 cal yr BP (Bennett and Fuller,

2002). These dates fall either directly on top of or very close to their predicted depths based on the Minnewaska and Mohonk age models (Fig. 5).

The interval spanning the late Pleistocene through the early Holocene is too compressed in the Mohonk core to provide reliable chronological control for climatic events (see, for example, the interval 125–140 cm, which spans nearly 5000 cal yr BP; Fig. 6). For this reason, we use the dates from the Minnewaska core to delineate events prior to 8485 cal yr BP. Both cores are used to define the chronology between 8485 and 4950 cal yr BP, and the late Holocene chronology is based on the dates in the Mohonk core since this time interval was not recovered in the Minnewaska core.

Dates of climatic transitions are determined from linear interpolation between adjacent radiocarbon dated and pollen stratigraphic intervals and are rounded to the nearest century in recognition of the low resolution of both cores. The fact that all dates are in stratigraphic order and show typical compaction trends with depth and the fact that there is no evidence of depositional hiatuses in either lake validates this approach. Given the highly variable stratigraphy of the Mohonk core in particular, however, we are aware of the fact that sedimentation rates may have varied and that the timing of climatic events may be somewhat different from the chronology presented here. Confirmation of the events described below awaits collection of higher resolution records for the area.

5.2. Late-glacial and Holocene vegetation and climate variability

The evidence contained in the Sky Lakes cores reveals substantial environmental change since the retreat of the Laurentide ice sheet. Pollen stratigraphy in both Minnewaska and Mohonk appears to reflect changes similar to those found throughout the Northeast and first described by Deevey (1939) and Davis (1969). While long known to reflect variations in temperature and precipitation, short-term shifts in pollen abundance may also be caused by responses to disturbance and successional dynamics (Davis, 1976; Maenza-Gmelch, 1997a; Oswald et al., 2007). As a result, reconstruction of past climatic conditions benefits from additional information from proxies such as lithology, sediment chemistry, and macrofossils. The climatic intervals presented below and depicted in Figs. 3, 7, and 9 draw on the totality of the evidence and demonstrate several wetter and dryer phases during the Holocene along with changes in temperature. The proxy interpretations used to reconstruct these climatic intervals are summarized in Table 2. The cycles uncovered show similarities and differences to events recorded in other northeastern lakes.

5.2.1. Period of deglaciation and late Pleistocene forest arrival (>12,900 cal yr BP)

The basal sediments in the Minnewaska and Mohonk cores (Figs. 2, 6) reveal that both records extend into the deglacial period in the mid-Hudson Valley region of New York. The layered sands and silts in the Minnewaska core suggest seasonal sedimentation and flow velocities sufficient to mobilize sand. Accordingly, it appears that Lake Minnewaska received meltwater inflows from patches of ice remaining in the landscape as the ridge was deglaciated. As the meltwater source disappeared and stream flow to the lake ceased, sedimentation abruptly changed from sand to clay. Fine-grained deposition is consistent with wind or sheetwash transport of glacial rock flour from the thin and patchy till deposits within the small catchment.

Coring did not penetrate a sufficient thickness of the basal sand unit at Mohonk Lake to determine whether it is laminated, but the abrupt transition from medium sand to clay indicates a similar history as at Minnewaska. The absence of plant macrofossils, low organic carbon content, and high pine pollen abundance of Moh-1 (Fig. 6) suggests that this zone reflects long-distance transport of pollen from outside the watershed prior to the local appearance of forest (Ibe, 1985; Peteet et al., 1993).

With the arrival of vegetation in the Minnewaska watershed, organic deposition increased (rising %C at 140 cm), and clastic inputs declined, probably as slopes were stabilized and thin till deposits became mantled in a protective layer of organic litter (Fig. 3). Zones Min-1 and Moh-2 show abundant *Picea* and *Abies* pollen along with *Pinus*, *Quercus*, and *Betula*, and together with zone Moh-1, are equivalent to the A1–A3 New England pollen zones of Deevey (1939) and Davis (1969) that reflect cool late glacial conditions recognized as the Bølling-Allerød warming. Conifer (Minnewaska), *Betula glandulosa* (Minnewaska), and *Picea* (Mohonk) macrofossils in zones Min-1 and Moh-2 (Figs. 4, 8), along with moderately abundant oak pollen (15% in Minnewaska and as much as 20% in

Mohonk), indicate a mixed thermophilous deciduous-boreal forest and a correspondingly cool and humid climate such as that found by Peteet et al. (1993) at Alpine Swamp, New Jersey, Linsley Pond, Connecticut and Allamuchy Pond, New Jersey (40 km south, 130 km east-southeast, and 105 km south-southwest of the Shawangunk ridge respectively), by Maenza-Gmelch (1997c) in the Hudson Highlands of New York (Sutherland Pond, 40 km south), and by Robinson et al. (2005) in the "black dirt" region of Orange County, New York (45 km southwest).

Expansion of forest in the Minnewaska watershed led not only to the deposition of plant macrofossils and an increase in organic carbon content, but also to a 7‰ shift in carbon isotopic composition toward more negative values (Fig. 3). $\delta^{13}C$ and C/N values show that Lake Minnewaska and Mohonk organic sediments are a mixture of terrestrial C_3 plants, aquatic bryophytes, and lacustrine algae (Meyers, 1994). While C_3 plants and lacustrine algae have identical $\delta^{13}C$ ranges, Galster (2001) and Lord (2003) found that aquatic algae from lakes in Vermont, New Hampshire, and Maine typically show more negative carbon isotopic values than the surrounding terrestrial vegetation. Based on this finding, they interpreted shifts toward more negative $\delta^{13}C$ values at the onset of post-glacial organic deposition to a gradual ramping up of aquatic ecosystems; increased forest cover led to higher nutrient influxes to lakes, thereby stimulating lacustrine production and causing the isotopic shift. While we did not measure isotopic values on the terrestrial and aquatic vegetation in the Minnewaska and Mohonk watersheds, we find Galster and Lord's mechanism to be a plausible explanation for the isotopic shift at Minnewaska in light of the extremely low nutrient levels of the Shawangunk conglomerate and the oligotrophic nature of the lake. Part of the negative isotopic shift is likely also related to the presence of the aquatic bryophyte *Fontinalis* (zone Min-2, Fig. 4), since aquatic bryophytes commonly display highly negative $\delta^{13}C$ values related to uptake of respired carbon dioxide (Proctor et al., 1992; Pentecost, 2000).
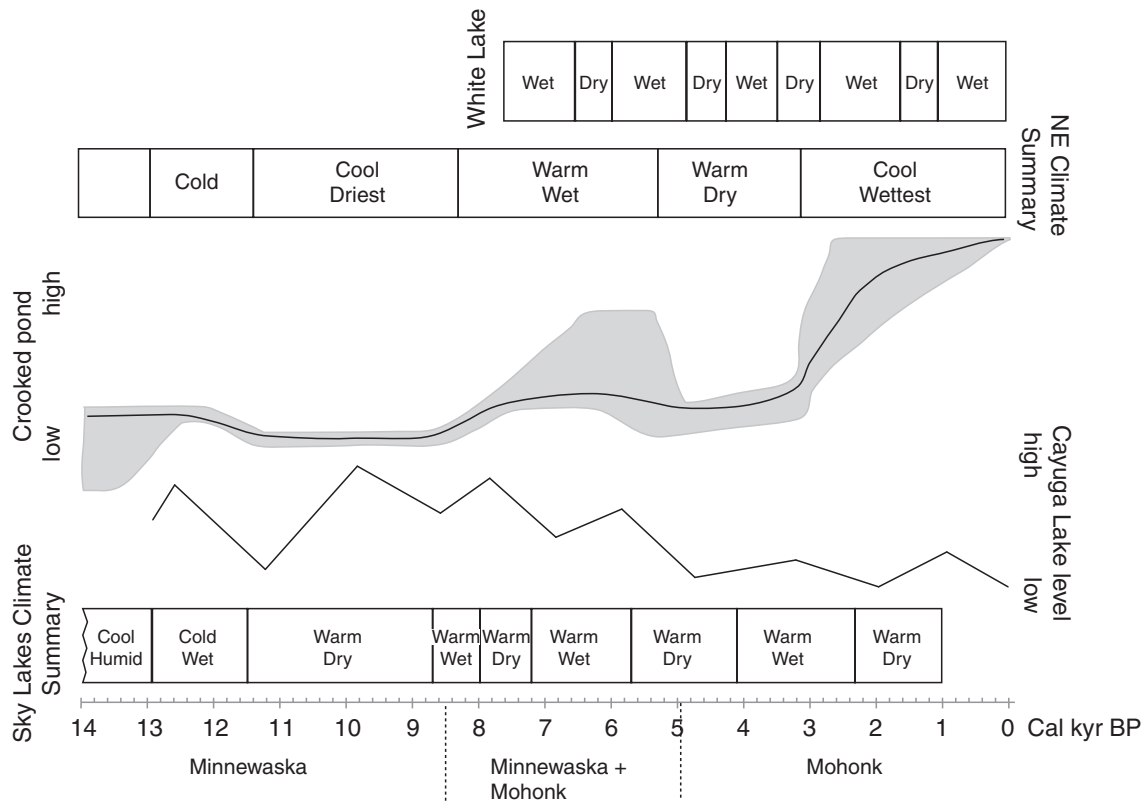

Fig. 9. Climate summary for the Sky Lakes record in comparison to the lake level history of Crooked Pond and New England climate summary (Shuman et al., 2004), the Lake Cayuga lake level history (Mullins, 1998), and wet–dry cycles at White Lake, New Jersey (Li et al., 2007). Gray shading around the Crooked Pond curve exhibits potential range of lake levels at any given time interval. Climate of the Shawangunk ridge alternates between matching that of New England and that of central New York, but refined chronologies are needed.

Table 2
Proxy interpretations.

Temperature	
Cold	*Warm*
Picea pollen and macrofossils	*Tsuga* pollen and macrofossils
Abies pollen	*Quercus* pollen
	Carya pollen

Precipitation	
Wet	*Dry*
Charcoal lacking	Charcoal abundant
High organic carbon	Low organic carbon
Tsuga pollen	*Castanea* pollen
Fagus pollen	*Pinus rigida* macrofossils
Alnus pollen	
Carya pollen	
Silty clay	

Water depth	
Shallow	*Deep*
Najas macrofossils	Lack of shallow water taxa
Chara macrofossils	
Nitella macrofossils	
Bryozoan statoblasts	

Disturbance	
High	*Low*
Sand	Gyttja
Alnus pollen	
Castanea pollen	
Vitis seeds	
Rubus seeds	
Compositae seeds	

Despite its proximity to Lake Minnewaska, Mohonk Lake shows a somewhat different post-glacial history. Sedimentology, C/N ratio, and δ^{13}C values show that organic deposition commenced with lacustrine algae (Fig. 7) followed by the appearance of the *Fontinalis* bryophyte, which produced a thick mat between 144 and 132 cm in the core. The presence of this bryophyte indicates that lake waters were clear and oligotrophic and thus able to support colonization on the floor or rocky sides of the lake.

5.2.2. Younger Dryas cooling (12,900–11,500 cal yr BP)

Relatively high amounts of *Alnus* and *Betula* pollen between 12,900 and 11,500 cal yr BP (125–110 cm depth) in the Minnewaska core (zone Min-2, equivalent to Deevey-Davis zone A4) reflect the Younger Dryas cooling event (Peteet et al., 1990; Mayle et al., 1993; Maenza-Gmelch, 1997c; Robinson et al., 2005). This event is not as readily apparent in the Mohonk core, though one sample at ~130 cm depth (zone Moh-3) shows elevated *Alnus* pollen and higher amounts of *Picea* than in zone Moh-2, suggestive of cooling (Peteet et al., 1990). Because *Alnus* and *Betula* are shade-intolerant, their increase during the Younger Dryas has been interpreted to reflect enhanced winter storm activity that caused overstory damage and allowed more light to reach the forest floor (Peteet et al., 1990, first documentation of Younger Dryas in the region). The colder, snowier conditions led to a peak in organic carbon content at ~11,700 cal yr BP in Lake Minnewaska, perhaps due to reduced oxidation of organic matter. This is in contrast to the record of organic carbon fluctuations at Crooked Pond and Makepeace Cedar Swamp in New England (~290 km east of the Shawangunk ridge), where organic carbon content declined during the Younger Dryas, attributed to a reduction in aquatic productivity (Shuman et al., 2004), but consistent with records in central New York that show increased organic matter deposition related to

increased nutrient influx to lakes associated with higher precipitation rates (Ellis et al., 2004).

In Mohonk Lake, the Younger Dryas was marked by a cessation of bryophyte growth (Fig. 6). The onset of silt deposition at 132 cm suggests a possible cause, which is an increase in turbidity that diminished photosynthetically active radiation levels at the lake floor. The silt further suggests increased inflow to Mohonk Lake from the ephemeral streams along its southern shore and therefore wetter conditions and/or typical increase in clastic sedimentation associated with colder conditions. In contrast, Davis, Rocky, and New Long Ponds in Massachusetts show lowstands from 13.4 to 10.9 cal kyr BP (Newby et al., 2011), suggesting drier conditions during the Allerød and the Younger Dryas. With the cessation of bryophyte growth and increased clastic inputs at Mohonk Lake, organic carbon content of the sediments fell, and organic inputs began to reflect the same mixture of aquatic algal and terrestrial inputs seen at Minnewaska (Fig. 7).

Despite the cold conditions associated with the Younger Dryas, Lake Minnewaska reveals the presence of *Tsuga* macrofossils as early as 12,700 cal yr BP (Fig. 4) similar to the arrival of *Tsuga* in sites such as Alpine Swamp, New Jersey (Peteet et al., 1990), Allamuchy Pond (Peteet et al., 1993), and Sutherland Pond, New York (Maenza-Gmelch, 1997c), but not as early as at Linsley Pond, Connecticut (Peteet et al., 1993) where it is evident in the Allerød interval as well as the second half of the Younger Dryas.

5.2.3. Onset of Holocene warmth and drought (11,500–8700 cal yr BP)

Zone Min-3, represented by only one sample, shows a return to high *Pinus* pollen percentages following their decline in zone Min-2 (Fig. 2), and represents the Deevey–Davis B zone in which *Pinus strobus* (Eastern white pine) dominated the record throughout the northeastern US (Deevey, 1939; Davis, 1969; Peteet et al., 1990, 1993; Maenza-Gmelch, 1997a; Toney et al., 2003; Robinson et al., 2005; Oswald et al., 2007). This zone is not apparent at Mohonk Lake, probably because of the low resolution of the core in this time interval. We did not attempt to quantify the abundance of *P. strobus* pollen relative to other *Pinus* species, but it is apparent in the Sky Lakes samples from this time interval. While there is no evidence of *P. strobus* needles in either Minnewaska or Mohonk during this time, abundant conifer leaf scales are present, presumably from this taxon. In addition to the higher *Pinus* pollen abundances, higher *Quercus* pollen and lack of *Fagus* and *Carya* pollen indicate that conditions became warmer and drier than during the preceding Younger Dryas time (Deevey, 1939; Davis, 1969; Peteet et al., 1990; Shuman et al., 2004; Oswald et al., 2007, 2010), a result consistent with evidence for lowstands in Davis, Rocky, and New Long Ponds in southern Massachusetts (Newby et al., 2011). In contrast, central New York experienced colder and wetter conditions from 10,000 to 9400 cal yr BP inferred from multiple proxies, including carbonate and total organic matter content and stable isotopes of carbonates (Ellis et al., 2004; McFadden et al., 2005).

The major expansions in *Quercus* and *Tsuga* pollen and the corresponding declines in *Pinus*, *Picea*, and *Abies* pollen that mark zones Min-4 and Moh-4 are equivalent to the C1 zone in the New York, New Jersey and New England pollen records (Deevey, 1939; Davis, 1969) and represent the further development of Holocene warmth (Figs. 2, 6). *Alnus* and *Betula* show steep declines as *Pinus* and *Quercus* replace these early successional species. A large decline in percent organic carbon in Lake Minnewaska reflects greater oxidation of organic sediments, and declining C/N ratio suggests a shift toward higher algal productivity, both consistent with warmer and drier conditions (Fig. 3). In Mohonk Lake, organic carbon also steeply declines, and carbon isotopic values undergo a positive shift, reflecting a change from the bryophyte deposition of late glacial time to mixed algal and C$_3$ vegetation (Fig. 7). Charcoal is present in Lake Minnewaska, and first appears in the Mohonk record, increasing toward the top of zone Moh-4 (Fig. 8).

Drier conditions on the Shawangunk ridge occurred simultaneously with drier climate in southeastern New York and New Jersey as well

as in southern New England. In the nearby Hudson Highlands (Maenza-Gmelch, 1997a) drought is indicated by minimal values of loss on ignition (LOI) and higher charcoal influx rates. At Alpine Swamp, the C1 oak-dominated pollen zone shows low LOI and abundant macrofossils from shallow water emergents such as *Decodon* (water-willow), *Cephalanthus* (buttonbush), *Scirpus* (bulrush), and *Dulichium* (Peteet et al., 1990). Similar dry conditions are revealed in Allamuchy Pond by the presence of *Decodon*, *Typha* (cattail), and *Eleocharis* (spikerush) (Peteet et al., 1993). At Ballston Lake near Saratoga, NY, ~135 km to the north of the Shawangunk ridge, LOI reveals minimal values from about 10,000–7000 cal yr BP, with minor oscillations (Toney et al., 2003). Lithologic shifts at Makepeace Cedar Swamp and Crooked Pond in eastern Massachusetts, at Echo Lake in New Hampshire (Newby et al., 2000; Shuman et al., 2004; Fig. 9), and at lakes in southern Ontario (Haas and McAndrews, 2000) demonstrate how widespread drought conditions were. In contrast, marls deposited in wetlands fringing Cayuga Lake, 235 km northwest of the Shawangunk ridge, indicate higher than modern lake levels produced by a wetter climate (Mullins, 1998). Consistent with these findings, carbonate proxies in a core from the eastern basin of Lake Ontario suggest warm and wet conditions between 9400 and 5300 cal yr BP (McFadden et al., 2005).

5.2.4. Early Holocene wet interval (8700–8000 cal yr BP)

A rise in *Tsuga*, *Fagus*, and *Betula* pollen, a rise in organic carbon content, and a decline in *Quercus* pollen between 8700 and 8000 cal yr BP in the Minnewaska core (zone Min-5) shows a change to wetter conditions (Fig. 2). Peak wetness dates to 8100 cal yr BP (7260 ^{14}C yr BP) and may reflect the 8.2-ka event seen in Greenland ice cores (Alley et al., 1997). This event appears to have brought colder, drier, and windier conditions to Greenland (Alley et al., 1997), the Great Lakes region (McFadden et al., 2005), and to parts of New England (Newby et al., 2009, 2011), while Owasco and Cayuga Lakes in central New York indicate higher than modern lake levels (Dwyer et al., 1996; Mullins, 1998). In other New York–New Jersey sites a short-lived wet interval is not obvious, so an alternative explanation for the Minnewaska pollen and carbon content changes could be that cooler conditions led to lower evaporative stress and to the appearance of wetter conditions in an otherwise drier climate. Dating rapid events may also be problematic in defining this interval.

Whatever the conditions, this interval appears to have lasted longer in the Sky Lakes than in other sites, though we note that this may be an artifact of chronological control. The long duration of this event is consistent with results at Blood Pond, Massachusetts, however, where Hou et al. (2006) found a cold event lasting from ~9000 to 8000 cal yr BP that they correlated to the 8.2-ka event; they also cautioned that chronological control was insufficient to ensure a 1000-year duration for that event.

5.2.5. Early Holocene warmth and drought (7900–7300 cal yr BP)

The poorly sorted sand horizon between 7900 and 7300 cal yr BP (119 and 112 cm) in the Mohonk core indicates an episode of landscape instability in which soil erosion increased. Charcoal records (Fig. 8) reveal that fire played an important role in this period of instability, suggesting that it was caused by drought rather than by increased storm activity as was interpreted for similar inorganic layers in Ritterbush Pond, Vermont (Brown et al., 2002). Corroborating evidence for drier conditions comes from Lake Minnewaska, which shows an interval of high *Quercus* pollen, lower *Tsuga* and slightly lower *Fagus* pollen, and low organic carbon deposition from 7900 to 7300 cal yr BP (67–52 cm, zone Min-6, Figs. 2, 3).

Drier conditions at this time are not supported by lake level studies in New England, where climate appears to have been mostly wet between ~8000 and 5500 cal yr BP (Newby et al., 2000; Shuman et al., 2004; Fig. 9). Exceptions to this generally wetter climate appear in the Davis Pond record in southwestern Massachusetts, which shows brief lowstands at 7.7 and 6.6–6.4 cal kyr BP during this interval (Newby et al., 2011). Furthermore, drier conditions are inferred in

central New York, where Lake Cayuga experienced a small lake level drop during an interval of wetter than modern climate (Mullins, 1998). Elsewhere in the Northeast, oscillations in LOI at Alpine Swamp suggest possible moisture fluctuations during this time, with lower LOI values prior to 7000 cal yr BP (Peteet et al., 1990).

We suggest that the extreme sensitivity of the Sky Lakes watersheds to drought, as discussed in Section 2, may have allowed Lakes Mohonk and Minnewaska to record a negative change in water balance too small to trigger a response in many of the New England records. Alternatively, differences in records may reflect regional variations in climate or problems with dating dry intervals. Because the landscape instability that characterizes drought events often contributes a mixture of sand, silt, and older soils to lakes, the dating of drier intervals can be too old in lake records and may preclude reliable correlations.

5.2.6. Middle Holocene wet interval (7100–5700 cal yr BP)

Declining *Quercus* pollen, rising *Tsuga* pollen, low *Pinus* pollen, increasing organic carbon content, and rising C/N ratio in Lake Minnewaska (48–15 cm, pollen zone Min-7) reveal a change toward wetter conditions generally consistent with the evidence in New England (Shuman et al., 2004, 2005; Oswald et al., 2007; Newby et al., 2011; Figs. 2, 3). In addition, a small increase in *Carya* pollen at 7000 cal yr BP (44 cm) suggests warmer conditions, and may correlate to the onset of New England pollen zone C2 (Deevey, 1939), dated to 7070–7200 cal yr BP (6180–6230 ^{14}C yr BP) in the nearby Hudson Highlands (Maenza-Gmelch, 1997a). In Mohonk Lake, charcoal concentrations fall when compared to the early Holocene drought interval (see Moh-4 to Moh-5 boundary in Fig. 8), and sediments change from the poorly sorted sand deposited during the drought to silt and then clayey gyttja, both consistent with more mesic conditions.

While the onset of the middle Holocene wet interval is the same in the two Sky Lakes cores (7290 cal yr BP in Minnewaska and 7270 cal yr BP in Mohonk), the end of the interval appears to differ. In Mohonk, the sedimentology suggests wet conditions ended by 6400 cal yr BP (see Moh-5 to Moh-6 boundary in Fig. 7), whereas in Minnewaska wet climate continued through 5700 cal yr BP (see Min-7 to Min-8 boundary in Fig. 3). The end of the mid-Holocene mesic interval in the Mohonk core is marked by an abrupt transition to poorly sorted sand (102 cm, Fig. 6) that was deposited during a period of landscape instability caused by drought (see below). Accordingly, the mismatch in ages for the close of the mesic conditions may be due to compaction of the clayey gyttja layer in the Mohonk core by the denser sand, which would lead to an apparent earlier age for the end of the wet interval. Alternatively, or perhaps additionally, part of the Mohonk record may have been lost to scour of the gyttja by the incoming sand. Given these issues, we use the Lake Minnewaska date to define the end of the interval and the onset of mid-Holocene drought.

In central New York, the timing of wet conditions differs somewhat from the Sky Lakes and New England chronologies. Lake Ontario proxies indicate increased throughput of water between 6800 and 5000 cal yr BP (McFadden et al., 2005), and Cayuga Lake experienced a highstand centered about 5800 cal yr BP (Mullins, 1998).

5.2.7. Middle Holocene drought and hemlock decline (5700–4100 cal yr BP)

The poorly sorted sand horizon between 102 and 87.5 cm and the overlying sandy gyttja from 87.5–78 cm in the Mohonk core contain charcoal and macrofossils of *Pinus rigida*, a fire adapted pine (Seischab and Bernard, 1996) not found in any other part of the record (zone Moh-6 in Fig. 8). In addition, elevated concentrations of *Najas* seeds, *Chara*, and *Nitella* suggest that lake level dropped. *Najas flexilis* is a rooted aquatic plant found in warm shallow water while *Chara* and *Nitella* are floating aquatic algae commonly found in shallow waterbodies (Haas and McAndrews, 2000). Elevated concentrations of bryozoan statoblasts also indicate lower water levels since bryozoans are most commonly found in shallow littoral zones (Crisman et al., 1986). Elevated C/N ratios

show a shift toward terrestrial vegetation inputs within this interval, which, when combined with the evidence already presented, suggests increased soil erosion caused by frequent fires. The presence of *Vitis*, *Rubus*, and Compositae seeds, all species favoring open conditions, high light, and poorer soils, confirm the magnitude of the drought disturbance.

At Minnewaska, mid-Holocene drought is marked by a decline in organic carbon content, an increase in *Pinus* pollen that may reflect increased pitch pine across the Shawangunk ridge, an increase in *Quercus* pollen, and a decline in *Tsuga* pollen (zone Min-8 in Fig. 2). Large quantities of *Tsuga* branchlets deposited at the onset of zone Min-8 (Fig. 4) suggest a species in collapse, a finding mirrored at Mohonk Lake where *Tsuga* pollen drops by a half to two thirds of its previous abundance in the middle of zone Moh-6 (Fig. 6). The onset of *Tsuga* collapse in Minnewaska, based on linear interpolation between the 4380 and 5530 ^{14}C yr BP dates, occurs at 5420 cal yr BP (4770 ^{14}C yr BP), in very good agreement with the 5520 cal yr BP (4750 ^{14}C yr BP) regional hemlock decline date of Bennett and Fuller (2002) and the 5300 cal yr BP date at Ballston Lake, Saratoga, NY (Toney et al., 2003). Accordingly, we assign the 5520 cal yr BP date to the middle of zone Moh-6 (see Figs. 5–8), assuming that the *Tsuga* decline was synchronous in the Sky lakes. Hemlock decline appears to have occurred later in the nearby Hudson Highlands, beginning near 4700 cal yr BP (4180 ^{14}C yr BP) at Sutherland Pond and near 3800 cal yr BP (3540 ^{14}C yr BP) at Spruce Pond (Maenza-Gmelch, 1997a).

While the organic carbon record from Lake Minnewaska indicates that the mid-Holocene drought began around 5800 cal yr BP, pollen and macrofossil evidence shows that it was most intense at 5500 cal yr BP, the time of highest charcoal, bryozoan statoblast, *Najas* seed, *Chara*, and *Nitella* concentrations in the Mohonk core and of highest *Tsuga* branchlet concentration in the Minnewaska core (Figs. 4, 8). These findings support the notion that drought and then fire played an important role in the mid-Holocene collapse of *Tsuga* (Miller, 1973; Haas and McAndrews, 2000; Foster et al., 2006), at one time attributed to an arboreal pathogen outbreak (Davis, 1981). *Tsuga* has very shallow rooting systems and its range is limited westward by soils and atmospheric moisture (Fowells, 1965). It is also very sensitive to fire (Foster and Zebryk, 1993). The similarity of the Sky Lakes records of drought to the sequence of LOI decline followed by a sharp drop in *Tsuga* seen at Alpine Swamp (Peteet et al., 1990) is paralleled in the Hudson Highlands by a sharp drop in LOI in Spruce Pond at least 800 years prior to the *Tsuga* decline (Maenza-Gmelch, 1997a). Likewise, as at Sutherland Pond (Maenza-Gmelch, 1997a), declines in *Tsuga* are coeval with maximal charcoal accumulation rates for the entire Holocene. Recovery of *Tsuga* on the Shawangunk ridge (beginning of zone Moh-7) appears to have commenced around 4100 cal yr BP following a return to more mesic conditions (below), consistent with the onset of recovery throughout New England (Foster et al., 2006) but more than a millennium earlier than in the Hudson Highlands (Maenza-Gmelch, 1997a).

The intense drought conditions and die-off of *Tsuga* during the middle Holocene may have contributed to an expansion of *Castanea* on the Shawangunk ridge, as previously proposed for the northeastern US (Miller, 1973; Foster and Zebryk, 1993). While this tree became more abundant in New England as conditions became cooler and somewhat wetter following the mid-Holocene drought (after 3000 cal yr BP, Deevey, 1939; Shuman et al., 2004), both Mohonk and Minnewaska show major increases in pollen percentage (zones Min-8 and Moh-6 in Figs. 2, 6) and total flux (not shown) coinciding with drought and *Tsuga* decline. *Castanea* is known to be drought tolerant but shade intolerant (Thompson et al., 1999; Bauerle et al., 2006; McEwan et al., 2006; USDA-NRCS, 2010), so its expansion during *Tsuga* decline may have been a result of the opening of canopy gaps that provided more light to the forest floor. It is also known to be a prolific sprouter and to thrive in disturbed areas (Foster and Zebryk, 1993). Its peak abundance at 25% pollen in the Mohonk core is remarkable for the region as it is generally underrepresented in the pollen rain (Paillet et al., 1991), and is probably

related to the dry, thin soils on the Shawangunk ridge as it is favored by these conditions (Miller, 1973).

Drought conditions appear to have extended from 5600 to 3200 cal yr BP at Makepeace Cedar Swamp (Newby et al., 2000) and from 5400 to 3000 cal yr BP at Crooked Pond and Echo Lake (Shuman et al., 2004), nearly a millennium longer than the Sky Lakes record indicates (Fig. 9). In comparison, Davis Pond shows a pronounced lowstand from 5600 to 4900 cal yr BP (Newby et al., 2011) and Cayuga Lake shows a dramatic drop in water level between ~5800 and 4750 cal yr BP, the latter followed by rising water levels up to ~3200 cal yr BP (Mullins, 1998), though levels never reach values as high as those in the early to middle Holocene.

The variability in sediment compressibility and low number of radiocarbon dates in the Mohonk core makes it difficult to correlate the Sky Lakes record to the New England and central New York records with confidence, and some of these latter records also do not have well-constrained chronologies. Macrofossil records show declining charcoal, bryozoan statoblast, and *Najas* seed concentrations toward the end of zone Moh-6, suggesting increasing effective wetness. In contrast, a variety of proxies in the eastern basin of Lake Ontario suggest cooler and drier conditions between 5300 cal yr BP and A.D. 1850 (McFadden et al., 2005).

5.2.8. Late Holocene wet interval (4100–2300 cal yr BP)

Wetter conditions appear to have prevailed on the Shawangunk ridge between 4100 and 2300 cal yr BP based on the disappearance of bryozoan statoblasts, *Najas*, and *Chara*, a large reduction in charcoal, and a return to clayey gyttja sedimentation in the Mohonk core (zone Moh-8 in Figs. 7 and 8). Wetter conditions are also supported by recovery of *Tsuga* and expansion of *Fagus*. Many northeastern US records indicate substantial cooling in the late Holocene (Deevey, 1939; Davis, 1981; Shuman et al., 2004), which may have contributed to increased effective wetness on the Shawangunk ridge. In contrast, Davis Pond shows substantial lowstands at 3.5 cal kyr BP and from 3.0 to 2.3 cal kyr BP (Newby et al., 2011). Similarly, pollen evidence from Wildwood Lake, Long Island, New York indicates declining *Fagus* between ~4000 and 2000 cal yr BP, suggesting drier conditions. Oswald et al. (2010) note that a shift in seasonality of precipitation toward drier summers might have led to the Wildwood Lake *Fagus* decline despite higher annual precipitation receipts, as originally proposed by Shuman and Donnelly (2006). Such a mechanism is not supported on the Shawangunk ridge where *Fagus* increased at the same time.

5.2.9. Late Holocene drying (2300–1000 cal yr BP)

A final episode of drying on the Shawangunk ridge between 2300 and 1000 cal yr BP is suggested by the reappearance of bryozoans and increased charcoal in the Mohonk core (zone Moh-8, Fig. 8). *Najas* seed concentrations remain very low, and neither *Chara* nor *Nitella* reappear, suggesting that this interval of drier conditions was less intense than the middle Holocene drought. Indeed, the change in lithology from gyttja to silt at 11 cm, combined with a drop in bryozoan statoblast abundance and lower charcoal values, suggests that conditions became more humid toward the end of the dry period, beginning around 1400 cal yr BP.

While Mohonk Lake recorded this late Holocene drying episode, many lakes in New England appear to have reached maximal levels between 3000 and 0 cal yr BP (Shuman et al., 2004; Fig. 9). In contrast, Cayuga Lake fell between 3200 and ~2000 cal yr BP, after which it rose until ~950 cal yr BP (Mullins, 1998). The center of the Cayuga drought correlates generally with the dry interval seen at Mohonk, and the lake level rise seen at Cayuga may correlate to the increased humidity seen toward the end of the late Holocene drought.

In sum, the Holocene records from Lakes Minnewaska and Mohonk contain lithology, pollen, and macrofossil sequences that show both similarities and differences to inferred wet–dry shifts in various parts of the Northeast. Collection of higher resolution sediment cores with multi-proxy information from the greater Catskill

region is warranted to better constrain regional climatic variability, particularly intervals of drought.

5.3. Causes of northeastern climate variability

A number of factors may be responsible for the wet–dry cycles recorded in the Sky Lakes cores. Webb et al. (1993) used relationships between modern pollen distributions and values of precipitation, mean January and July temperatures, and soil moisture to interpret changes in fossil pollen records at 3000 year time slices from the present through late glacial time. They demonstrated climatic changes consistent with ice sheet retreat and variations in orbital forcing. While these factors certainly contributed to the climatic changes recorded in the Sky Lakes cores on the Shawangunk ridge, the abrupt changes in moisture balance apparent in these lakes require additional explanation.

Burnett et al. (2004) identified four primary synoptic-scale patterns influencing winter precipitation in central New York, including 1) a "lake effect" pattern characterized by low pressure over New England and high pressure over the upper Midwestern states that drives northwesterly winds, 2) a "warm sector" pattern in which winds from the south and southwest blow toward a low pressure zone centered over eastern New York, 3) a "coastal" pattern in which a low pressure zone moves northward along the East Coast drawing air in from the east, and 4) a "warm frontal overrunning" pattern in which low pressure centered over western Kentucky and Tennessee drives northward moving warm fronts. Summer precipitation is similarly complicated and depends on the position of the polar jet stream (Mullins et al., 2011).

Given the myriad influences on northeastern climate, drought and wet intervals might be expected to reflect atmospheric circulation patterns such as the El Niño Southern Oscillation (ENSO), North Atlantic Oscillation (NAO), and Pacific North American pattern (PNA). Bradbury et al. (2002) regressed various climatic variables against indices for all of these patterns and demonstrated weak, but statistically significant, correlations between the NAO and modern stream flow in New England. Low flow, which they took to indicate drought conditions, coincided with cold North Atlantic sea surface temperatures and a negative phase NAO that caused an eastward shift of the East Coast trough. This shift carried frontal storms associated with the polar front jet away from land and out to sea, leading to the drier conditions. These results would seem to support the interpretations of Li et al. (2007), who attributed repeated Holocene drought intervals in White Lake, New Jersey with similar timing to events in Davis Pond, Massachusetts (Newby et al., 2011) to millennial scale ice rafting events in the North Atlantic and noted the similarity of the relationship between drought and colder sea surface temperatures during the Holocene to the modern negative phase of the NAO.

A regional climate model simulation for the Northeast provides additional information on modes of climatic variability, showing that the combination of a negative phase NAO and positive phase PNA generates drier than average winters in eastern New York and New England, but wetter than average winters in the Great Lakes region. The latter are caused by enhanced lake effect snows, a result confirmed by meteorological observations (Notaro et al., 2006). Enhanced lake effect snows can also result from climatological warming in the Great Lakes region as demonstrated for the years 1931–2001 by Burnett et al. (2003). These findings may help to explain some of the contrasting behavior throughout the Holocene between lakes in central New York and New England, in particular the high lake levels exhibited by Lakes Cayuga and Owasco during times of lowstands at Crooked Pond, Makepeace Cedar Swamp, and Echo Lake. The Sky Lakes, falling geographically between these regions, might be expected to show conditions intermediate or alternating between them, and indeed, the Minnewaska and Mohonk records at times show more affinity to central New York and at other times more similarities to New England.

The strong mid-Holocene drought that affected the Shawangunk ridge between 5800 and 4100 cal yr BP coincides with widespread drought conditions in the western interior and Midwest (Menking and Anderson, 2003). Examining three late 19th century droughts that extended from coast to coast in the conterminous U.S., Herweijer et al. (2006) found that persistent La Niña conditions contributed to aridity. Similarly, in weather records from the 20th century, Griffiths and Bradley (2007) found that a combination of La Niñas and negative phase NAOs produce an increased number of consecutive dry days throughout southern New England, the southern tiers of Vermont, New Hampshire, and Pennsylvania, and the eastern third of New York while bringing a reduced number of consecutive dry days to central New York and northern Pennsylvania. Thus a combination of enhanced La Niñas and colder North Atlantic sea surface temperatures during the middle Holocene may be responsible for the drought intervals recorded by the Sky Lakes and may help explain why Cayuga and Owasco Lakes were often out of phase with New England.

Yin (2005) provides another potential explanation for drought conditions. In 15 different climate models used to simulate 21st century climate under conditions of increased atmospheric pCO_2, storm tracks shifted poleward and the northern and southern annual modes (NAM and SAM) switched toward their high index values. A significant shift of storm tracks northward would result in decreased precipitation in the Northeast, and may have occurred in previous droughts, as warming associated with insolation changes altered atmospheric circulation patterns. Indeed, Kirby et al. (2002a,b) used such a mechanism to explain oxygen isotopic shifts in Fayetteville Green Lake in central New York throughout the Holocene. These workers demonstrated a strong modern correlation between mean polar vortex position over the area and isotopic composition and abundance of winter precipitation, showing that during periods of vortex contraction, snow and rain originate primarily from western, continental sources, which tend to be isotopically depleted and relatively lacking in moisture. In contrast, periods of vortex expansion allow air masses from the southeast to bring a greater abundance of more enriched precipitation to the region. The oxygen isotopic composition of lacustrine marls deposited in Fayetteville Green Lake changed throughout the Holocene toward more depleted values, indicating polar vortex contraction in response to orbital precession-driven increases in winter insolation and suggesting a cause of the mid- to late-Holocene decline in lake levels seen in Lakes Owasco, Cayuga, and Canandaigua (Kirby et al., 2002a,b).

The low resolution of the Sky Lakes record and limited age control discourage further speculation on the particular causes of any individual drought episode on the Shawangunk ridge. However, the record leaves cause for concern regarding future water resources. A return to prolonged regional drought conditions such as occurred at various times during the Holocene, including the Medieval Warm Period drought documented in the lower Hudson Valley (Pederson et al., 2005), would severely threaten water supplies for the New York City metropolitan region, making it imperative that a more detailed drought investigation is conducted in the Catskill Mountain region.

6. Conclusions

Evidence from the Minnewaska and Mohonk cores provides the following conclusions to the questions posed in the introduction of this paper:

1) Climate appears to have oscillated between wetter and drier conditions repeatedly throughout the Holocene. Oscillations show both similarities and differences to New England and central New York, reflecting the geographically intermediate location of the Sky Lakes.
2) Like New England but in contrast to the Finger lakes in central New York, the Shawangunk Ridge appears to have experienced dry

conditions at the onset of the Holocene that gave way to wetter conditions shortly after 8700 cal yr BP.

3) Declining lake levels as determined from Lake Mohonk plant macrofossils, an increase in charcoal abundance, and appearance of *Pinus rigida* in the middle Holocene pre-date the *Tsuga* decline, suggesting that prolonged drought acted as a triggering mechanism for species collapse.

4) Drought episodes may reflect a combination of enhanced La Niña conditions due to different orbital forcing during the early to middle Holocene, negative phase NAO patterns associated with cold North Atlantic sea surface temperatures, and positive phase PNA conditions. Northward shifts in storm tracks associated with Holocene warming provide another possible mechanism for the production of extended droughts.

5) Prolonged droughts are a cause of concern with regards to the water resources of the New York City metropolitan region.

Acknowledgments

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