Geochronology and geomorphology of the Jones Point glacial landform in Lower Hudson Valley (New York): Insight into deglaciation processes since the Last Glacial Maximum

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A B S T R A C T

The glacial deposits at Jones Point, located on the western side of the lower Hudson River, New York, were investigated with geologic, geophysical, remote sensing and optically stimulated luminescence (OSL) dating methods to build an interpretation of landform origin, formation and timing. OSL dates on eight samples of quartz sand, seven single-aliquot, and one single-grain of quartz yield an age range of 14–27 ka for the proglacial and glaciofluvial deposits at Jones Point. Optical age results suggest that Jones Point deposits largely predate the glacial Lake Albany drainage erosional flood episode in the Hudson River Valley ca. 15–13 ka. Based on this data, we conclude that this major erosional event mostly removed valley fill deposits, leaving elevated terraces during deglaciation at the end of the Last Glacial Maximum (LGM). Our analysis supports the earlier 15 ka date for the flood occurrence and suggests that more recent overlying glaciofluvial deposits formed in the lee-side of the Jones Point promontory.

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1. Introduction

The lower Hudson River Valley in New York state has a well-known glacial history beginning with work by Merrill (1891) and continuing into the twenty-first century (Peet, 1904a, 1904b; Fuller, 1914; Thwaites, 1946; Newman et al., 1969; Markl, 1971; Merguerian and Sanders, 1990, 1992; Sanders and Merguerian, 1998; Cadwell et al., 2003). The evidence for multiple glacial episodes during the Pleistocene Epoch (Fuller, 1914) was supported by studies of Sanders and Merguerian (1998). The major glacial advance and retreat of the Laurentide Ice Sheet (LIS) in North America left substantial late Quaternary sedimentary records in mid-lower Hudson River Valley, New York, partially reworked and eroded by sudden meltwater discharge (Uchupi et al., 2001; Donnelly et al., 2005; Rayburn et al., 2005; Stanford, 2010).

The maximum ice extent in this region is limited to the south by the Ronkonkoma moraine (Fuller, 1914) on Long Island (Fig. 1), and the upper age limit for the late Wisconsinan glacial maximum (LGM) is 21,750 ± 750 YBP, a 14C date based on bulk material of unfossiliferous silt from the moraine (Sirkin and Stuckenrath, 1980; Cadwell and Dineen, 1987; Muller and Calkin, 1993). More recent synthesis of local studies of glacial deposits and landforms in New Jersey, Long Island, and the Hudson valley, revision of the routing of the Hudson River, and the effect of the moraine’s breaches and sea level rise on the history of glacial lakes was done by Stanford (2010). This study in addition to varve chronology by Ridge (2004) for the northeastern region of the USA, place deposition of the terminal moraine (Ronkonkoma) between 21,500 and 20,000 YBP, confirming dates previously reported by Sirkin and Stuckenrath (1980). The details of within-valley deglaciation processes following terminal moraine breach and climate change causing northward shrinkage of the LIS are less constrained. Nonetheless, they are critically important to understanding the development of soils, revegetation, and ecological successions that eventually supported agriculture and subsistence for first settlers.

One of the earliest and widely used interpretations of deglaciation in New York State, including mid-lower Hudson River Valley, was summarized by Connally and Sirkin (1986). In their model of northward glacial retreat, a series of ice-dammed lakes and associated glaciofluvial deposits such as deltas, kames, moraines, and terraces were deposited. In addition, Oszvath (1985) and Cadwell et al. (2003) emphasized the effect of bedrock topography and relief on the formation of glacial features and their distribution: this helped to explain ice flows and glacial lake formation along the Hudson River. These interpretations are still being used in recent publications on deglaciation in New York, including research on the role of the freshwater discharge from glacial lakes into...
the Atlantic shelf (Uchupi et al., 2001; Donnelly et al., 2005; Rayburn et al., 2005; Stanford, 2010).

According to Connally and Sirkin (1986), northward LIS retreat and melting behind the Ronkonkoma moraine formed proglacial Lake Hudson, which was bounded to the north by the Shenandoah/Pellets Island Moraine (Fig. 1). The Harbor Hill moraine, located to the north of Ronkonkoma moraine on Long Island (Fig. 1), is a result of the late Wisconsinan ice advance (Cadwell et al., 2003). Connally and Sirkin also proposed that the glacier had receded to the Shenandoah/Pellets Island Moraine position by 17.95 cal. ka BP and a subsequent advance (Rosendale) took place about 16.1 cal. ka BP. After 16.1 cal. ka BP the ice retreat continued northward, providing space for southern-derived ecological succession of plants and animals.

With continued glacial retreat, proglacial Lake Hudson that occupied areas of the lower Hudson (Fig. 1) connected with proglacial Lake Albany (stretched from the Mid-Hudson to the north); features and landforms associated with this process are ice contact deltas from tributaries that mark paleolake levels (Connally and Sirkin, 1986). Sanders and Merguerian (1998) reported evidence of loess deposition (however, no geochronological data is available) at Croton Point (Fig. 1), an important indicator of ice sheet absence at this location. Other loess locations were recorded in Long Island (de Laguna, 1963) but also not dated. Ridge (2004) compiled varve data from Antevs (1928) and Stanford and Harper (1991) to establish Haverstraw and Hackensack deglaciation margins in the lower Hudson River Valley (Fig. 1).

Timing and interpretation of the deglaciation process is also important to understanding evolution of species in the lower Hudson Valley. For example, separate fish studies on incipient speciation in subwatershed tributaries demonstrated the existence of genetic changes in four freshwater fish species (Arcement and Rachlin, 1976; Miller, 2015), which is attributed to their isolation by saltwater intrusion from the Atlantic Ocean. Because of the limited age control on glacial deposits in the mid-lower Hudson Valley, such research currently lacks a detailed chronological time frame in which these events occurred.

Geomorphic and detailed sedimentary studies using limited bulk radiocarbon samples in the northeastern US have contributed to understanding and interpretation of existing glacial features. Cosmogenic radionuclide surface exposure (10Be) and 14C dating techniques helped establish better constraints on its deglaciation history. Surface exposure dating using 10Be by Balco and Schaefer (2006) on boulders from moraines in eastern Connecticut, in conjunction with the previously published New England varve chronology, place deglaciation within a range of 18,500–19,000 YBP. In southern New York, boulder ages from Bear Mountain State Park area (Fig. 1) range from 18,000 to 22,000 YBP (Schaefer pers. comm.). Peteet et al. (2012) used new accelerated mass spectrometer (AMS) 14C dates from 13 sites, including one near Jones Point in the area of Harriman State Park (Fig. 1), that show a range of deposits dating to 15–16 cal. ka BP. The discrepancy in 10Be and 14C geochronologies led Peteet et al. (2012) to propose a hypothesis of severe climatic conditions in North America that postponed complete deglaciation until 16 cal. ka BP, the oldest 14C date obtained in the region. This can be possibly associated with Rosendale readvance that took place ≈16,100 YBP (Connally and Sirkin, 1986).

In this paper, we report the first optically stimulated luminescence (OSL) dates of quartz sand and glacial sediments exposed in the lower Hudson River Valley near Jones Point, New York. Integration of the OSL data with geomorphic methods used in this study helped to identify landform morphology, structure, and evolution since LGM. Data incorporated from previous studies helped place the geomorphic evolution of the studied landform in a regional context of major deglaciation events: catastrophic breach of the Narrows moraine releasing Lake Albany cold waters into the Atlantic and possible triggering of glacial readvance.

1.1. Luminescence dating of glacial sediments

The long continental deglaciation process produces variable environmental conditions and depositional features, some of which are better suited for luminescence dating than others (Rhodes, 2011; Wyslouch et al., 2015). Luminescence dating, described in more
detail later, is applied to sedimentary grains of quartz and feldspar to determine an age estimate of the last time sediment was exposed to sunlight or thermal exposure \(300 \, ^\circ\text{C}\) (Huntley et al., 1985).

After removal from the stimulation source (light/heat), the mineral dosimeters acquire and retain a dose of environmental radiation equivalent to time spent in burial and subject to surrounding radioactive decay (Aitken, 1998).

Complete removal of a previous dose (known as bleaching) during sedimentary transport is ideal, and if not achieved, understanding the degree to which that previous dose has been removed is essential for accurate age determination (Alexanderson and Murray, 2007; Preusser et al., 2007). A review by Fuchs and Owen (2008) emphasized the importance of transport distance on dose resetting and related this to glacial deposits. They conclude that partial resetting is a likely scenario for glaciofluvial, glaciolacustrine, and glaciomarine deposits. Numerous independent studies have substantiated this (e.g., Glasser et al., 2006; Thomas et al., 2006; Lepper et al., 2007; Rittenour et al., 2015; King et al., 2014). These studies successfully applied luminescence dating techniques to Patagonian, Russian, North American and Norwegian glacial material (respectively) by taking into account detailed sedimentology and stratigraphy, comparison to \(^{14}\text{C}\) ages, use of a minimum age model (Galbraith and Roberts, 2012), single-grain dating, and/or examination of residual luminescence signals in modern glacially-derived sediments.

The effect of transport distance sedimentary history on luminescence dating of glacial sediments is twofold, bleachability (already mentioned) and sensitivity (Rhodes, 2011). Sensitivity is described by the OSL intensity, and a lack of luminescence intensity (or ‘dim’ signals) has been identified in glacial quartz from the New Zealand Alps (Preusser et al., 2006). The authors attributed these low detectable quartz OSL signals to short transport history and lack of luminescence traps. The notion of sensitization is likely derived from sedimentary cycling, which exposes grains to radiation over time and increases the number of available luminescence traps (Sawakuchi et al., 2011). Another noteworthy observation for glacial quartz is the lab-induced increase in sensitivity and thermal transfer that can occur with repeated laboratory dosing and stimulations (Rhodes, 2000; Preusser et al., 2006). The OSL ages may be overestimated if left uncorrected for such a shift in sensitivity or if it is too dominant to correct for (Rhodes, 2000).

To summarize, sediments farthest from the glacial front or supraglacial sediments typically have a better chance of luminescence signals being completely reset prior to burial than those deposited subglacially, intraglacially, or proximal to the glacial front, though this is somewhat complicated by bedrock lithology; this closely links OSL dating and its interpretation with deglaciation processes and landforms. In this paper, we combine geochronologic data from OSL dating and from geologic and geophysical methods in order to describe in detail the glacial depositional landform at Jones Point, NY, and place it in the framework of deglaciation history of the region.

1.2. Study area: Jones Point, New York

Two exposures near Bear Mountain State Park at Jones Point (Fig. 1) are the focus for this study. Following a landslide in 2001, site JP_2001 was exposed on the flank of the slope next to New York State route 9W (Cusick and Gorokhovich, 2003) (Fig. 2). Due west across route 9W is site JP_2002, located in an abandoned gravel quarry (Fig. 3). Currently, both sites are heavily modified by surface erosion and slumping. These deposits are juxtaposed on the steep granite and gneissic bedrock (New York State Geologic Map, 1:250,000) slope of Jones Point where the Hudson River is channeled through an abrupt

Fig. 2. Glacial landform (JP_2001) exposed after the landslide in 2001, facing southwest. Route 9W passes on the left; letters refer to sedimentary deposits sampled for granulometric analysis; layers D, G, I, and J were sampled for OSL (see Table 1 for sedimentology).

Fig. 3. Repeat photograph of glacial landform JP_2002 in 2002 and 2016. Exposure is south facing, route 9W passes on the right, the bedrock of Jones Point is on the left behind the trees; upper layer of boulders and gravel (layer K) was considered as a marker in the two locations. Two samples for OSL were taken here in 2016, one above and one below layer K that dips in western direction.
bend around Dunderberg Mountain (Fig. 4). The first descriptions of glacial deposits at Jones Point can be found in Merrill (1891) and Peet (1904a, 1904b). Both authors described deposits as a terrace, less than a mile long, composed of stratified subrounded gravel, sand, and little clay. The original deposit morphology was cut through at the end of the nineteenth century by the construction of route 9W; however, no records or documentation were available from New York State Department of Transportation.

2. Methodology

To study the timing and deglaciation environments at Jones Point, NY, we employed OSL dating, sedimentary and petrographic analysis, electrical resistivity geophysics, and remote sensing (LiDAR). Sedimentary texture and sorting were used to identify suitable layers for OSL dating and to determine the onset of fluvial activity during glacial retreat. Additionally, the boulder/gravel layer K was analyzed by petrographic analysis of thin sections to determine the source (local vs. distant) of its material. Bare-earth (i.e., without vegetation cover) LiDAR data were used to investigate the overall morphology of the landform. Electrical resistivity was employed to conduct a subsurface investigation of the landforms and to verify the continuity of boulder layer K between our two sites.

2.1. Optically stimulated luminescence (OSL) dating

Luminescence dating can be applied to sedimentary quartz and feldspar in a variety of Quaternary geomorphic deposits (Aitken, 1998). Upon burial or removal from the heat source, these minerals are capable of storing a measurable dose of radiation in which they become natural dosimeters (Huntley et al., 1985). While OSL is complementary to radiocarbon and cosmogenic nuclide dating in terms of timescale over which it is applied (late Quaternary), OSL dating requires no organic material, is minimally influenced by subsequent erosion or aggradation, and it can date widespread surficial geologic material beyond the limit of radiocarbon and below the typical limit with cosmogenic exposure dating (Duller, 2000).

Standard OSL dating of quartz sand utilizes blue-green wavelengths to release excited electrons back to their natural energy orbital (Huntley et al., 1985). Upon ejection from geologically stable traps, some electrons recombine in luminescence centers to produce a measurable amount of light. These photons make up the luminescence signal measured in a dark room laboratory and are calibrated through irradiation cycles in order to calculate an equivalent dose $D_E$ in units of gray (Gy, where 1Gy = Joule/kg) of radiation (Murray and Wintle, 2000). Electrons become ionized and trapped in mineral defects through continued exposure to nuclear radiation within the surrounding sediment (Aitken, 1989). The decay of this radiation is converted to a dose rate $D_R$ in this case through chemical analysis such as inductively coupled mass spectrometry (ICP-MS) and known conversion factors (Adamiec and Aitken, 1998; Aitken, 1998; Guérin et al., 2011).

The OSL age (ka) is the quotient of the $D_E$ (Gy) and $D_R$ (Gy/ka). One main assumption for OSL dating is that the material sampled was thoroughly exposed to sunlight or high heat to reset, or zero out, any and all previously acquired dose prior to the most recent burial or event of interest. If mineral grains are not thoroughly zeroed, they will retain an antecedent dose (Murray et al., 1995), a phenomena known as partial bleaching or incomplete zeroing. If left unaccounted for, partially bleached material will overestimate the latest burial dose and age of an OSL sample (i.e., Olley et al., 1999). This problem can be amplified in glacial settings as transport distances may be short and transport energy and sediment load typically are high and/or occur subglacially (Fuchs and Owen, 2008; Wyschnytsky et al., 2015). Methods are employed to mitigate partial bleaching effects in these OSL samples, and those are discussed in the Results section.

2.1.1. OSL sample collection

The OSL samples were collected in 2014 and 2016 from two exposures of a continuous outcrop of interbedded sand and pebble/cobble/boulder stratigraphy at site JP_2002 (n 2) and Site JP_2001 (n 6) at Jones Point, NY.
NY (Figs. 2, 3). Thick sand beds and lenses (>30 cm) were the main targets for OSL sampling to avoid DR heterogeneity; however, some thinner sand lenses between pebble/cobble/boulder units were sampled given the glacial context that often lacks abundant sand deposits. Layers with original sedimentary structures, i.e., planar laminations, were sought after to minimize post-depositional mixing and/or high energy and rapid deposition, though massive sands were also sampled. Layers with original sedimentary structures, i.e., planar laminations, were sought after to minimize post-depositional mixing and/or high energy and rapid deposition, though massive sands were also sampled. Metal conduits cut into 8-inch pieces were hammered horizontally into targeted layers for DE samples (Fig. 5). The DR sediment was uniformly collected within a 15-cm radius of the DE sample; cobbles and gravels were included where they intersected the DR sphere of influence.

2.1.2. OSL sample processing

Eight OSL samples were processed and analyzed at the Utah State University Luminescence Laboratory in Logan, UT, USA, for small-aliquot and single-grain single-aliquot regenerative dose (SAR) analysis of quartz sand (Duller et al., 1999; Murray and Wintle, 2000). Samples were opened under subdued amber safe lighting (~590 nm) and wet sieved to a grain size range between 125 and 250 μm. The quartz mineral fraction was isolated by using 10% hydrochloric acid and bleached to dissolve carbonates and organic material, sodium polytungstate (2.7 g/cm³) to separate out the heavy minerals, and concentrated hydrofluoric and hydrochloric acids to remove the feldspars, etch the outer quartz grain, and prevent formation of fluorite precipitates (see Wintle, 1997 for details). Aliquots were checked for feldspar contamination using infrared (IR) stimulation on all aliquots following the SAR protocol. Single-grains were checked for feldspar following the OSL IR depletion ratio protocol of Duller (2003).

2.1.3. Dose rate (DR) determination

Representative subsamples of bulk DR material were uniformly divided into ~30 g samples and sent to the ALS Minerals Laboratory in Elko, Nevada, to measure the radioelemental concentrations of K, Rb, Th, and U using ICP-MS and ICP-AES analyses. In situ gravimetric moisture content was measured at the USU Luminescence Lab on all samples. Dose rate calculations include cosmic contribution by using sample

Fig. 6. Proposed source areas I, II, and Catskills Mountains. These areas potentially contributed material to layer K at Jones Point. More detailed description of rock lithologies within each source area are in Tables 1 and 2.
depth, elevation, and longitude/latitude following Prescott and Hutton (1994), influence of water attenuation (Aitken and Xie, 1990), uncertainty in elemental measurements, and dose rate conversion factors (Guérin et al., 2011).

2.1.4. Optical measurements

Individual aliquot $D_E$ s on quartz were calculated using the SAR technique of Murray and Wintle (2000) on 1–2 mm diameter small-aliquots (~10 grains per disk) for seven out of the eight samples. Optical measurements were performed on Risø TL/OSL Model DA-20 readers with blue-green light-emitting diodes (LED) (470 ± 30 nm) as the stimulation source. The luminescence signal was measured through 7.5-mm UV filters (U-340) over 40–45 s (250 channels) at 125 °C with LED diodes at 70–90% power (~45 mW/cm²), and calculated by subtracting the average of the last 5 s (background signal) from the first 0.7 s of the signal decay curve. Aliquots were heated to >180 °C prior to all optical measurements to remove the thermally unstable traps around the 110 °C TL peak in quartz (Rhodes and Bailey, 1997). Preheat temperatures at 240 °C (held for 10 s) followed natural and regenerative beta doses, and test dose preheat treatments were 160–220 °C (held for 10 s). Two samples (USU-1765 and USU-1766) required a high temperature (280 °C) optical stimulation (bleach) with blue LEDs held for 40 s at the end of each cycle to clear out high levels of recuperation (Wintle and Murray, 2006). Dose response curves were fit within linear, saturating-exponential or saturating-exponential plus linear fits to calculate individual aliquot $D_E$ values.

Single-grain dating was applied to USU-1768 and analyzed using the SAR technique (Duller et al., 1999; Murray and Wintle, 2000) on a Risø TL/OSL Model DA-20 reader with single-grain attachment (Duller et al., 1999; Bøtter-Jensen et al., 2003). Grains were heated to 125 °C then stimulated with a green laser (532 nm) at 90% power (45 W/cm²) for 0.8 s with a 0.1 s pause before and after stimulation. The luminescence signal was detected through a 7.5-mm UV filter (U-340). Following natural and regenerative doses, the preheat temperature was 240 °C and held for 10 s, and test dose preheat treatments were 220 °C also held for 10 s. Dose response curves were fit within saturating-exponential and saturating-exponential plus linear fits to calculate individual grain $D_E$ values.

2.2. Electrical resistivity geophysics

To investigate the three-dimensional morphology of the terrace stratigraphy at JP_2001 (Fig. 2), a geophysical survey was designed using the electrical resistivity method, which is commonly used to distinguish coarser (more resistive) from finer (less resistive) clastic sediments. The survey was conducted using an Agi Supersting 28-electrode DC resistivity unit, with two-dimensional inverse modeling of the subsurface, and included two perpendicular profiles.
along the ridge behind the bluff exposure and across a saddle to another ridge overlooking the Hudson (Fig. 4). With a 3-m electrode spacing, apparent resistivity pseudosections were corrected for topography and generated using the Schlumberger and Dipole-dipole configurations, and the data were inverted using EarthImager2D (Advanced Geosciences Inc., Austin, TX) to produce subsurface distributions of apparent resistivity for interpretation. Data were not excessively noisy, and standard inversion error criteria such as RMS and normalized L2 were low (<3% and 2.4 respectively).

2.3. Remote sensing (LiDAR) application

LiDAR (light detection and ranging) is widely used in remote sensing for mapping earth surfaces and the shallow subsurface (Carter et al., 2001; Slob and Hack, 2004). Vertical accuracy of measurements depends on data interpolation and horizontal displacement and can vary between 17 and 26 cm (Hodgson and Bresnahan, 2004). American Society for Photogrammetry and Remote Sensing (ASPRS) has standard categories on LiDAR vertical accuracy from 5 cm (quality level 0) to 18.5 cm (Quality Level 3) (USGS, 2014).

LiDAR data for this study were obtained from U.S. Geological Survey (USGS) data application EarthExplorer (see: http://earthexplorer.usgs.gov/) in the American Society for Photogrammetry and Remote Sensing (ASPRS) LAS format that stores three-dimensional point cloud data and associated attributes. A LiDAR LAS file usually contains information on signal returns (e.g., first, second, third, fourth, last, etc.) and its classification (e.g., unassigned, ground, noise, water, reserved, etc.) with a number of associated points in the point cloud. To map the surface of the bare earth, we used ground class points and created an image of Jones Point with bare earth elevation, i.e., without vegetation cover (Fig. 4). This helped to visualize the terrain as a geomorphologic entity and interpret the glacial landform.

Fig. 8. Glaciofluvial features of the site JP_2001: cross-bedding (layer C), horizontal layers (E), and meltwater deposits of layer F; blue arrows indicate location of these features on the landform exposure. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Fig. 9. An example of the sample and corresponding thin section of the sample from layer K.
2.4. Sedimentary and petrographic analyses

Field descriptions of sedimentary units exposed at JP_2001 and JP_2002 include color, texture, sedimentary structure, and contact appearance.

Twelve layers were targeted at site JP_2001 for granulometric analysis using freely available software package GRADISTAT (Blott and Pye, 2001) developed for Microsoft Excel.

Layer K is a coarse-grained sedimentary marker bed containing small boulders and is observed at similar elevations at both sites (~25–31 m asl, Figs. 2, 3). Both locations were mapped with high-precision differential correction GPS (Trimble Pro). We targeted 12 boulders (>10 cm diameter) of unique lithology and had thin sections prepared for petrographic analysis with a petrographic microscope to identify minerals and rock types. Using the 1:250,000 digital geologic map (Fisher et al., 1970) in geographic information systems (GIS) (Dicken et al., 2005), we grouped lithologies and identified three main source areas (source areas I, II, and Catskills Mountains, Fig. 6) to which we matched the rock types.

3. Results

3.1. Sedimentary (granulometric) analysis

Detailed sedimentary analysis was done only for the site JP_2001 because JP_2002 was not accessible at the time. Table 1 includes results of field descriptions on unit thickness, sedimentary structure, and granulometric analysis conducted with GRADISTAT; we used Folk and Ward (1957) graphic measures for mean, median, and sorting.

Mean particle size of sediments ranges from 0.04 to 4.82 mm. One sample (layer F) was primarily silt (0.04 mm), seven samples represent fine to coarse sand, and two upper samples represent fine gravel. Sorting in all layers is moderate (≥1.62) with only two samples exhibiting moderate sorting: layers J and D. Layer H has sorting coefficient 1.7 on the borderline, close to 1.62.

Fig. 7 shows distribution of main granulometric parameters with depth, from layers A to J. Median diameter in all layers follows closely mean size distribution. Layers A and B exhibit poor sorting and large mean/median grain size. Improvement of sorting coefficient and corresponding reduction in mean/median grain size is noticeable in layers C, D, E. Farther down, layer F shows finest mean/median grain size but poor sorting, the pattern similar to layer I. Other layers (G, H, I, and J) show that increase in grain size corresponds to decrease in sediment sorting (i.e., increase in sorting coefficient).

Specific sedimentary structures such as plane beds and cross-bedding were observed in sedimentary layers C, D, and E. Layer F represents a thin fine-grained lens, possibly left by rapid meltwater. Fig. 8 shows examples of these features documented at site JP_2001.

Granulometric analysis and stratigraphy of sedimentary units helped to group 12 studied layers in three main facies (Table 1): glacial drift (layers A–B), glaciofluvial (C–F) and proglacial deposits (G–J and L). Glacial drift in top layers A and B is a matrix-supported diamict, containing rounded and subrounded gravel material, poorly sorted. Glaciofluvial layers C–F contain cross-bedding structures; whereas the thickest basal layer L shows massive, coarse, planar beds.

3.2. Petrographic analysis

Analysis of thin sections made from rock samples (Fig. 9) found in the JP_2002 site showed a list of rock types (Table 2) that were compared to source areas I and II and to the Catskills Mountains. Table 3 shows unique occurrences of rock types within source areas I, II, and Catskills Mountains.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Identified rock type</th>
<th>Possible source area</th>
</tr>
</thead>
<tbody>
<tr>
<td>JP1-1</td>
<td>Quartz-rich granitoid</td>
<td>Source area I</td>
</tr>
<tr>
<td>JP1-2</td>
<td>Metamorphic quartzite</td>
<td>Source area I</td>
</tr>
<tr>
<td>JP1-3</td>
<td>Low grade metamorphic quartzite with carbon grains/calcite veins</td>
<td>Source area I (?)</td>
</tr>
<tr>
<td>JP1-4</td>
<td>Sedimentary metamorphic sandstone</td>
<td>Source area II</td>
</tr>
<tr>
<td>JP1-5</td>
<td>Sedimentary metamorphic sandstone</td>
<td>Source area II</td>
</tr>
<tr>
<td>JP1-6</td>
<td>Sandstone</td>
<td>Source area II</td>
</tr>
<tr>
<td>JP1-7</td>
<td>Granite</td>
<td>Source area I</td>
</tr>
<tr>
<td>JP1-8</td>
<td>Fine-grained limestone</td>
<td>Catskills Mountains</td>
</tr>
<tr>
<td>JP1-9</td>
<td>Sandstone (with carbonate inclusions)</td>
<td>Catskills Mountains, source area II</td>
</tr>
<tr>
<td>JP1-10</td>
<td>Altered sericite granite</td>
<td>Source area I</td>
</tr>
<tr>
<td>JP1-11</td>
<td>Quartz monzonite breccia</td>
<td>Source area II</td>
</tr>
<tr>
<td>JP1-12</td>
<td>Monzosyenite</td>
<td>Source area II</td>
</tr>
</tbody>
</table>

3.3. Geophysical analysis

LiDAR data analysis was conducted to establish geomorphology of the area of study to verify potential connection between JP_2001 and JP_2002 and their relationship with the bedrock side of the valley. Fig. 4 shows rendering of the surface topography without vegetation cover. Route 9W is shown and cuts through the large terrace that was possibly one feature before construction and development of the area, mentioned by Merrill (1891) and Peet (1904a, 1904b). Lower right part of the image displays railroad tracks from the New York Hudson Line. Flat surfaces (terrace) are evident at locations 1, 2, 3, and 4.

3.4. Geophysical analysis

The resistivity surveys were conducted at sites 2 and 3 (Fig. 4), where LiDAR imagery shows remnant terraces of glaciofluvial materials extending east from the valley wall consisting of gneiss and diorite bedrock beyond route 9W. The profiles extended NNE-SSW across the terrace behind JP_2001 at site 2 (Fig. 10) and WNW-ESE perpendicular across the saddle to site 3 (Fig. 11). These orientations were designed to distinguish whether boulder horizons were laterally continuous and whether the horizon geometries were subhorizontal, draped, or steeply angled (Eaton et al., 2017).

The survey configurations (dipole-dipole and Schlumberger) at each location essentially show the same subsurface geometries, with high-resistivity materials (interpreted as the boulder horizon) mainly as
noncontinuous and limited to one end of the profile. In the case of the terrace remnant at site 2 closer to the valley wall (Fig. 4), the higher resistivity materials are at the NNE (left) end, and at site 3 they are found at the ESE (right) end of traverse 2 (Fig. 11). The depth of the high-resistivity materials (~3 m from the land surface) is consistent with the location of the boulder horizon where it appears in the southern bluff of site 2; however, no similar bluff outcrop was identified at site 3. Given the likely ice-proximal deglaciation scenario of the Hudson River Valley, it seems probable that the high-resistivity deposits have longitudinal axes parallel to the valley wall and the geophysical profiles crossed them at high angles if not perpendicularly in the case of site 3 (Fig. 11). Although the profiles appear subhorizontal at the site-scale, a comparison of similar boulder horizons at sites JP_2001 and JP_2002 suggests that the larger scale geometry is dipping in the downvalley direction (see Fig. 3). Therefore, the boulder horizon seems unlikely to be a planar horizontal topset bed as originally hypothesized (Eaton et al., 2017).

3.5. OSL dating analysis

In luminescence dating, aliquots or grains may be removed from further age analysis if they do not meet acceptance criteria relating to steps in the SAR protocol, such as repeated or no dose measurements or nonsaturating exponential luminescence signal growth with given doses. For the Jones Point OSL samples, individual aliquots were accepted in age calculation if they had no feldspar contamination, repeated dose ratio within 40% of unity (USU-1766 and USU-1767 omitted), recuperation at the zero dose SAR step <15% of the natural signal, and if the natural and test doses fell within the nonsaturating part of the dose response curve (Fig. 12). In addition, grains were rejected from analysis if the initial signal (0.1–0.13 s) was <3 times the background (0.67–0.90 s) after receiving a beta dose. Cumulative $D_E$ values were calculated using the central age model (CAM) or three-parameter minimum age model (MAM-3) of Galbraith and Roberts (2012) of at least nine acceptable aliquots or grains of quartz sand (Table 4).
Fig. 12. Luminescence dose-response curves (left) and OSL response (right) for two 1-mm small-aliquot examples (USU-2422 and USU-1765) and one single-grain example from USU-1768. Top example (USU-2422) is an exceptionally sensitive aliquot, while the middle example (USU-1765) is less sensitive to luminescence and more common for OSL samples presented here. The lower sensitivity and greater $D_E$ distribution has led to increased total error in OSL age estimates. Luminescence-responsive single grains were even less common, and the bottom example is one of a kind. The 675 out of 700 (96%) grains stimulated were not included in further analysis because initial response was <3 times background level.
The chosen age model was based on sedimentology and stratigraphic context, overdispersion parameter, and skewness of aliquot or grain distribution (Fig. 13; Table 5). All samples had overdispersion, or variance in the $D_d$ data, ≥40%, a sign that the distribution may contain some portion of partially bleached grains (i.e., Arnold et al., 2007). Wyschnytsky et al. (2015) had shown this to be a common scenario in glaciofluvial sediments and that high overdispersion values do not always warrant use of the MAM when considering sedimentology that suggests lower energy deposition and stratigraphic context such as thinner sand layers and discontinuous lenses. Table 5 lists OSL sample sedimentology, overdispersion, and skewness in stratigraphic order. All but three samples were calculated with the MAM (Galbraith and Roberts, 2012) and are in stratigraphic order given error limits.

Errors on $D_d$ and age estimates are reported at 2σ standard error and include errors related to instrument calibration; and dose rate and equivalent dose calculations were calculated in quadrature using the methods of Aitken and Aldred (1972) and Guérin et al. (2011). The largest contributor to OSL age error is that from the $D_d$ values, not surprising given the spread of individual measurements (Fig. 13). Error estimates on measured aliquots or grains were higher because of the lower peak signals (≤500 counts/s) at the small mask size (1-mm). Initial test aliquots were run at 2-mm mask, but many of these $D_d$s were likely overestimated from the presence of partially bleached grains. The smaller 1-mm mask size was chosen to mitigate issues with partial bleaching, though signal resolution was reduced. Single-grain dating is preferable for partially bleached glacial sediments (Duller, 2006; Thrasher et al., 2009); however, we were only moderately successful with single-grain dating on USU-1768 given the low signal response of these samples. Fig. 12 displays dose-response curves and optically stimulated luminescence decay from three samples. One aliquot from USU-2422 has an exceptionally bright fast component, error. While the example from USU-1765 was more commonly found, resulting in greater (worsened) recycling ratios, poorly fit dose-response curves and thus higher $D_d$s errors. These dimmer signals, in the example given ≤200 counts/s for a 78 Gy beta dose, are commonly found in sediments with shorter transport history and less time spent in the sedimentary cycle prior to burial (Preussner et al., 2006). While it would have been advantageous to apply single-grain dating to all samples in this study, given the poor dose response and large percentage of grains (675/700 = 96%) that were rejected on USU-1768 for low signal response, single-grain dating is not feasible for the samples.

Dose rate values range from 2.08 to 2.91 Gy/ka, and the lack of significant outliers in the chemistry data (Table 6) suggest little heterogeneity amongst the dose rate environment for the samples collected along this dual outcrop exposure. Moisture content values ranged from 1.6 (41.28° N, 73.95° W, 0.03 km) were used with burial depth to calculate cosmogenic contribution to the total dose rate following Prescott and Hutton (1994).

### 4. Discussion

In this work, we employed a variety of methods at Jones Point to better understand the geomorphic and sedimentary history of the glacial landform as it fits in the framework of the last deglaciation period. In terms of morphology, the LiDAR survey revealed a set of four distinguishable terraces (locations 1, 2, 3, and 4) along the flank of the Jones Point bedrock (Fig. 4) valley side. It is possible that construction of the railroad and a highway (route 9W) caused changes in their original shape. At the same time, flat surfaces and almost equal elevation levels in all four locations make it possible to conclude that it was indeed the large terrace described by Merrill (1891) and Peet (1904a, 1904b), previously identified as a kame terrace left behind by the downwasting process of deglaciation.

Studies by Oszvath (1985) and Cadwell et al. (2003) showed dependency of the deglaciation process in New York on topography of its bedrock. Cadwell et al. (2003) indicated that “recession of the ice margin was accompanied by thinning and regional downwasting thus uncovering upland summits” (Fig. 14). Later work (Moss, 2016; Stanford, 2010) showed that the Hudson River Valley has been deeply eroded by multiple glaciations and that sediments have been deposited and removed by multiple ice-dam glacial lake release episodes since the LGM. If the Jones Point sediments were deposited in contact with ice as it was retreating back up the narrow Hudson River Valley, one would expect more fine-grained glaciolacustrine sediments in this sequence of Hudson Valley deposits at this elevation (+25 m asl). Indeed, such sediments are likely to have been emplaced early but could have been eroded and removed if we assume that the Hudson River Valley became the major regional drain of the receding ice sheet, generating high volume meltwater flows down the constricted valley, removing glaciolacustrine and other fine sediments, and depositing and eroding existing proglacial massive sandy deposits (Fig. 14).

Sedimentary results (Moss, 2016) from deep borings for the new bridge (Haverstraw Ice Margin location, Fig. 1) over the Tappan Zee (a broad section of Hudson River Valley just south of Jones Point) provide new insight into the chronology of these events. The boundary between varved silt and clay glaciolacustrine deposits and overlying sands and gravels similar to Jones Point is at a considerable depth of ~60 m bsl (below sea level) underneath the modern river. Furthermore, these sands and gravels in turn have been eroded to a depth of ~39 m bsl where they are over lain by more modern estuarine organic sediments. The analysis of two $^{14}$C samples (bulk analysis) from the fragmentary organic matter at the base of these sediments, interpreted as a marine incursion, were dated to 10,790–12,598 YBP (Donnelly et al., 2005).

At Jones Point (Table 7) two major lithofacies are present: a more massive, plane-bedded facies attributed to proglacial processes

### Table 4

Optically stimulated luminescence (OSL) age information.

<table>
<thead>
<tr>
<th>Sample num.</th>
<th>USU num.</th>
<th>Grain size (μm)</th>
<th>Num. of aliquots or grains</th>
<th>$D_d$ (Gy/ka)</th>
<th>$D_d$ ± 2σ (Gy)</th>
<th>OSL age ± 2σ (ka)</th>
<th>Age model</th>
</tr>
</thead>
<tbody>
<tr>
<td>D</td>
<td>USU-1765</td>
<td>125–212</td>
<td>9 (26)</td>
<td>2.91 ± 0.14</td>
<td>41.4 ± 17.3</td>
<td>14.2 ± 6.1</td>
<td>CAM</td>
</tr>
<tr>
<td>G</td>
<td>USU-1766</td>
<td>150–250</td>
<td>10 (27)</td>
<td>2.81 ± 0.13</td>
<td>45.6 ± 15.2</td>
<td>16.2 ± 5.6</td>
<td>CAM</td>
</tr>
<tr>
<td>I</td>
<td>USU-1767</td>
<td>125–212</td>
<td>13 (25)</td>
<td>2.08 ± 0.14</td>
<td>46.6 ± 17.1</td>
<td>22.4 ± 8.6</td>
<td>CAM</td>
</tr>
<tr>
<td>J</td>
<td>USU-1768</td>
<td>125–250</td>
<td>12 (700)</td>
<td>2.20 ± 0.17</td>
<td>58.5 ± 23.6</td>
<td>26.6 ± 11.2</td>
<td>CAM</td>
</tr>
<tr>
<td>JP1</td>
<td>USU-2422</td>
<td>150–250</td>
<td>12 (30)</td>
<td>2.19 ± 0.10</td>
<td>55.3 ± 15.2</td>
<td>25.3 ± 7.4</td>
<td>CAM</td>
</tr>
<tr>
<td>JP2</td>
<td>USU-2423</td>
<td>150–250</td>
<td>12 (33)</td>
<td>2.53 ± 0.12</td>
<td>69.5 ± 12.8</td>
<td>27.3 ± 5.7</td>
<td>CAM</td>
</tr>
<tr>
<td>JP3</td>
<td>USU-2424</td>
<td>150–250</td>
<td>14 (32)</td>
<td>2.32 ± 0.11</td>
<td>39.7 ± 12.8</td>
<td>17.2 ± 5.8</td>
<td>CAM</td>
</tr>
<tr>
<td>JP4</td>
<td>USU-2425</td>
<td>125–250</td>
<td>17 (32)</td>
<td>2.53 ± 0.12</td>
<td>56.0 ± 18.6</td>
<td>22.2 ± 7.7</td>
<td>CAM</td>
</tr>
</tbody>
</table>

* Age analysis using the single-aliquot regenerative-dose procedure of Murray and Wintle (2000) on 1–2 mm small-aliquots of quartz sand unless otherwise noted. Number of aliquots used in age calculation and number of aliquots analyzed in parentheses.

b Equivalent dose ($D_d$) calculated using the central age model (CAM) or minimum age model (MAM) of Galbraith and Roberts (2012).

c Age analysis using the single-aliquot regenerative-dose procedure of Murray and Wintle (2000) on single-grains of quartz sand. Number of grains used in age calculation and number of grains analyzed in parentheses.
(G–J, L) and an upper secondary glaciofluvial facies showing cross-bedding (C–F). The dominance of the massive facies, the presence of the coarse boulder facies K, and the relative lack of finer sediments (clays or silts, characteristic of glaciolacustrine or ice-contact sediments) suggests winnowing by transport processes during deposition.

Fig. 13. Radial plots of equivalent dose ($D_E$) distributions on accepted aliquots or grains for eight OSL samples from Bear Mountain, NY. Skew values with * are significantly positive following Bailey and Arnold (2006).
Given that most of our OSL ages at Jones Point (Table 7) considerably predate this erosional event, the Jones Point deposits are likely to be all that remains from a more extensive valley-fill proglacial deposit (Lonne and Nemec, 2004; Winsemann et al., 2007). Although they differ on timing, Donnelly et al. (2005) and Stanford (2010) provide a mechanism for such an erosional event: the catastrophic glacial lake draining as a result of the breach in the terminal moraine at the Narrows in New York City. We interpret the boundary between Jones Point proglacial and glaciofluvial deposits as an indicator of this flood erosion event that, according to our OSL data, could possibly occur between 14 and 16 ka. This time interval is consistent with dates of 13 or 15 ka, according to conclusions of Donnelly et al. (2005) and Stanford (2010).

Layer K, within the massive proglacial sequence, is present at the JP_2001 and JP_2002 locations; it was considered a marker bed, probably derived from a lateral moraine related to the LGM period. The youngest OSL age of layer K containing a variety of material from distant sources, is most likely the remnant of a lateral moraine dating to the end of the LGM (according to OSL) from which the fines have been winnowed by subsequent massive meltwater discharge.

The formation of overlying glaciofluvial layers (F–A) is most likely associated with reworking of proglacial deposits following the flood erosional event described earlier. Bedforms within the upper part of the sequence and flow indicators from multiple directions suggest that the preservation of the Jones Point terraces was in part because of their location in the lee of the Dunderberg Mountain massif, while the same sediments were being eroded in the center of the valley (Fig. 14). Layers C (above) and E (below) have characteristic cross-bedding and planar bedding structures. Layer F is a silt, poorly sorted, that forms a thin (5 cm) layer between coarse sand layers, indicating sharp change in surrounding fluvial conditions, probably a localized flood by the meltwater. Layers G, H, I, and J maintain massive bedding with a decrease of mean/median grain size from J to I, making a typical graded bed sequence going up from layer K to I. Fig. 14 summarizes in four panels our proposed morphologic development at the site.

The OSL dating results show sequentially younger ages from the bottom of the terrace (layer J) to the upper glaciofluvial layer D providing a range between 27 and 14 ka. The youngest OSL age of layer D (Table 7) in this suite is 14.2 ± 6.1 ka. The OSL ages of layer G are 16.2 ± 5.6 and 17.2 ± 5.8 ka; three ages for the lower I layer and the J layer are 22.4 ± 8.6, 22.2 ± 7.7, 26.6 ± 11.2, and 27.3 ± 5.7 ka.

To place the OSL ages provided here into the overall deglaciation context, we compare them to other geochronologic data available in the vicinity of the study area. On a highland part of the lower Hudson

### Table 5

<table>
<thead>
<tr>
<th>USU num.</th>
<th>Depth (m)</th>
<th>Sedimentology (thickness cm)</th>
<th>Depositional environment</th>
<th>OD (%)</th>
<th>Skew</th>
<th>OSL age ± 2σ (ka)</th>
<th>Age model</th>
<th>Layer</th>
</tr>
</thead>
<tbody>
<tr>
<td>USU-1765</td>
<td>1.69</td>
<td>Sand, medium, well sorted (20 cm)</td>
<td>Glaciofluvial</td>
<td>56.8 ± 15.9</td>
<td>0.38</td>
<td>14.2 ± 6.1</td>
<td>CAM</td>
<td>D</td>
</tr>
<tr>
<td>USU-1766</td>
<td>2.7</td>
<td>Sand, coarse, well sorted (85 cm)</td>
<td>Proglacial melt-out</td>
<td>58.5 ± 15.4</td>
<td>0.82*</td>
<td>16.2 ± 5.6</td>
<td>CAM</td>
<td>G</td>
</tr>
<tr>
<td>USU-2424</td>
<td>3.5</td>
<td>Sand, coarse, well sorted (85 cm); massive sand</td>
<td>Proglacial melt-out</td>
<td>28.5 ± 8.3</td>
<td>0.75*</td>
<td>17.2 ± 5.8</td>
<td>CAM</td>
<td>G</td>
</tr>
<tr>
<td>USU-1767</td>
<td>4.18</td>
<td>Sand, fine, moderately sorted (20 cm)</td>
<td>Proglacial</td>
<td>45.3 ± 11.5</td>
<td>0.48</td>
<td>22.4 ± 8.6</td>
<td>CAM</td>
<td>I</td>
</tr>
<tr>
<td>USU-2425</td>
<td>5.5</td>
<td>Med-coarse sand above boulder/cobble/pebbles (30 cm)</td>
<td>Proglacial</td>
<td>32.8 ± 8.0</td>
<td>-0.22</td>
<td>22.2 ± 7.7</td>
<td>CAM</td>
<td>J</td>
</tr>
<tr>
<td>USU-1768 (5G)</td>
<td>4.38</td>
<td>Sand, medium, well sorted (95 cm)</td>
<td>Proglacial</td>
<td>25.5 ± 12.2</td>
<td>-0.03</td>
<td>26.6 ± 11.2</td>
<td>CAM</td>
<td>J</td>
</tr>
<tr>
<td>USU-2423#</td>
<td>4</td>
<td>Sand over boulder/cobble/pebbles</td>
<td>Proglacial</td>
<td>24.9 ± 7.9</td>
<td>0.54</td>
<td>27.3 ± 5.7</td>
<td>CAM</td>
<td>J</td>
</tr>
<tr>
<td>USU-2422#</td>
<td>5.3</td>
<td>Coarse-med sand lens below boulder/cobble/pebbles (32 cm)</td>
<td>Pre-LGM proglacial</td>
<td>41.6 ± 10.8</td>
<td>0.51</td>
<td>25.3 ± 7.4</td>
<td>CAM</td>
<td>L</td>
</tr>
</tbody>
</table>

*Overdispersion (OD) represents variance in D_E data beyond measurement uncertainties, OD > 20% may indicate significant scatter caused by depositional or post-depositional processes.

b Skew values with * are significantly positive following Bailey and Arnold (2006).

c Equivalent dose (D_E) calculated using the central age model (CAM) or minimum age model (MAM) of Galbraith and Roberts (2012).

d Layers listed in stratigraphic order; see Table 1 for granulometric details.

e Samples were taken at JP_2002 site.

### Table 6

<table>
<thead>
<tr>
<th>Sample num.</th>
<th>USU num.</th>
<th>In situ H2O (%)a</th>
<th>Depth (m)</th>
<th>K (%)b</th>
<th>Rb (ppm)b</th>
<th>Th (ppm)b</th>
<th>U (ppm)b</th>
<th>Cosmicc (Gy/ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>D</td>
<td>USU-1765</td>
<td>7.7</td>
<td>1.69</td>
<td>2.05 ± 0.05</td>
<td>79.8 ± 32</td>
<td>8.4 ± 0.8</td>
<td>1.5 ± 0.1</td>
<td>0.17 ± 0.02</td>
</tr>
<tr>
<td>G</td>
<td>USU-1766</td>
<td>7.7</td>
<td>2.70</td>
<td>2.04 ± 0.05</td>
<td>81.7 ± 33</td>
<td>7.6 ± 0.7</td>
<td>1.5 ± 0.1</td>
<td>0.15 ± 0.01</td>
</tr>
<tr>
<td>I</td>
<td>USU-1767</td>
<td>15.1</td>
<td>4.18</td>
<td>1.51 ± 0.04</td>
<td>66.4 ± 27</td>
<td>6.4 ± 0.6</td>
<td>1.5 ± 0.1</td>
<td>0.12 ± 0.01</td>
</tr>
<tr>
<td>J</td>
<td>USU-1768</td>
<td>19.2</td>
<td>4.38</td>
<td>1.62 ± 0.04</td>
<td>70.6 ± 28</td>
<td>7.8 ± 0.7</td>
<td>1.7 ± 0.1</td>
<td>0.12 ± 0.01</td>
</tr>
<tr>
<td>JP1#</td>
<td>USU-2422</td>
<td>2.2</td>
<td>5.3</td>
<td>1.56 ± 0.04</td>
<td>73.3 ± 29</td>
<td>6.1 ± 0.6</td>
<td>1.3 ± 0.1</td>
<td>0.11 ± 0.01</td>
</tr>
<tr>
<td>JP2#</td>
<td>USU-2423</td>
<td>1.6</td>
<td>4</td>
<td>1.92 ± 0.05</td>
<td>78.7 ± 32</td>
<td>6.3 ± 0.7</td>
<td>1.3 ± 0.1</td>
<td>0.13 ± 0.01</td>
</tr>
<tr>
<td>JP3</td>
<td>USU-2424</td>
<td>4.3</td>
<td>3.5</td>
<td>1.76 ± 0.04</td>
<td>75.8 ± 30</td>
<td>5.2 ± 0.5</td>
<td>1.2 ± 0.1</td>
<td>0.14 ± 0.01</td>
</tr>
<tr>
<td>JP4</td>
<td>USU-2425</td>
<td>5.3</td>
<td>5.5</td>
<td>1.85 ± 0.05</td>
<td>81.5 ± 33</td>
<td>6.9 ± 0.6</td>
<td>1.4 ± 0.1</td>
<td>0.11 ± 0.01</td>
</tr>
</tbody>
</table>

a Assumed average of all samples, 7.9 ± 2.4%, as moisture content over burial history for samples with <7%.

b Radioelemental concentrations determined by ALS Minerals (Elko, NV) using ICP-MS and ICP-AES techniques.

c Following Prescott and Hutton (1994).

d Samples were taken at JP_2002 site.
Valley in Harriman State Park (Fig. 1). Peteet et al. (2012) obtained four AMS-based \(^{14}\text{C}\) dates from bogs (Picea needles, twigs, clay) with a range of 14–15.1 cal. ka. According to those authors, these findings represent (the initial colonization of the landscape coincident with widespread climatic amelioration and ice-free condition). In the nearby Bear Mountain State Park (Fig. 1) surface exposure dates based on \(^{10}\text{Be}\) (Schaefer, pers. comm.) produced a range of 18–22 cal. ka. These data were obtained from large boulders on the ridges, indicating the time when ice retreat exposed the bedrock and erratic boulders to cosmogenic nuclide accumulation.

Two kilometers northwest from Jones Point, at Iona Island marsh, at a depth of 28 m below modern sea level, bulk \(^{14}\text{C}\) date based on twigs (oak, pine, spruce) were recovered from organic silt-sand interface (bore UH-1; Newman et al., 1969) and produced a date of 12,500 YBP (or 14,700 cal. YBP). Interestingly enough, Peteet et al. (2012) samples do not contain any sand or silt, indicating stable depositional conditions at sampling sites. It is possible that the UH-1 location was subject to more active glaciofluvial conditions.

The uncertainty of OSL results is in part associated with partial sunlight resetting as seen with the high dispersion and skew toward higher doses in most \(D\text{e}\) distributions, often a side effect of OSL when applied to glacial deposits. Adding to the uncertainty is the low sensitivity of the quartz grains to optical stimulation, noted in Fig. 12 and supported by the high amount of luminescence-dead single grains (\(\sim 96\%\)). These luminescence ages and characteristics combined with sedimentology and geomorphology support our interpretation of proglacial deposits having experienced some transport.

5. Conclusions

This study aimed to interpret and date glacial deposits found at Jones Point, New York, and place them in the framework of the deglaciation process on the lower Hudson River Valley. Data obtained from various techniques (OSL, sedimentary analysis, LiDAR, geophysics, and petrographic analysis) and synthesis of available information on the deglaciation process in the lower Hudson Valley allow us to make following conclusions:

- New OSL ages on proglacial massive sandy sediments at ~25 m asl from 16,000 to 27,000 YBP at Jones Point predate current estimates of a proglacial lake-draining flood episode between 13,000 and 15,000 YBP (Donnelly et al., 2005; Stanford, 2010). Our interpretation suggests that these sandy proglacial valley fill deposits were probably much more extensive but were eroded by proglacial lake drainage as shown by stratigraphy and ages of deep valley fill below Table 7

<table>
<thead>
<tr>
<th>Facies (layers)</th>
<th>Thickness (cm)</th>
<th>OSL age (ka)</th>
<th>Lithofacies code a</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glacial drift (A-B)</td>
<td>153</td>
<td>NA</td>
<td>Dcm</td>
</tr>
<tr>
<td>Glaciofluvial (C-F)</td>
<td>117</td>
<td>14.2 ± 6.1</td>
<td>Sh, Sp</td>
</tr>
<tr>
<td>Proglacial (G-J)</td>
<td>263</td>
<td>16.2 ± 5.6</td>
<td>Sm</td>
</tr>
<tr>
<td></td>
<td></td>
<td>17.2 ± 5.8</td>
<td></td>
</tr>
<tr>
<td></td>
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<td>22.4 ± 8.6</td>
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<td>22.2 ± 7.7</td>
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<td>26.6 ± 11.2</td>
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</tr>
<tr>
<td></td>
<td></td>
<td>27.3 ± 5.7</td>
<td></td>
</tr>
<tr>
<td>Lateral moraine remnant (K)</td>
<td>200</td>
<td>NA</td>
<td>Gcm</td>
</tr>
<tr>
<td>Proglacial (L)</td>
<td>&gt;300</td>
<td>25.3 ± 7.4</td>
<td>Sm</td>
</tr>
</tbody>
</table>

* Codes for lithofacies used from (Eyles et al., 1983; Benn and Evans, 1998).
sea level underneath the modern river. A boulder/cobble/pebble layer (K) embedded in these massive sands may be a remnant of lateral moraine with a distant source of rock material, most likely from the Catskills and Adirondack mountains.

- At Jones Point, the overlying glacioluvial sediments show bedforms indicating a quieter glacial environment starting ca. 14.2 ± 6.1 ka, sheltered in the lee of Dunderberg Mountain. Therefore, attribution of the lake-draining flood to the earlier 15 ka date (Stanford, 2010) is more suitable in the context of Jones Point OSL data. Moreover, this interpretation supports the theory of interum of the deglaciation process (Peteet et al., 2012) and Rosendale readvance (Connelly and Sirk, 1986) ca. 16 ka because massive cold flood water contribution to the Atlantic could likely cause local climatic fluctuations (Donnelly et al., 2005).

- The appearance of cross-bedding (layers C and D) coincides with continuous ice retreat from Jones Point by 14 cal. ka BP, as evident from 14C calibrated data by Peteet et al. (2012) and Newman et al. (1969); ice-free conditions and ensuing ecological succession triggered the formation of organic matter analyzed by 14C. A future OSL target in the area should be loess deposits described by Sanders and Merguerian (1998) at Croton Point.

- Results of this study allows reconstituting the following sequence of morphogenetic events at the Jones Point site: glacial advance during LGM-deposited lateral moraine (layer K); glacial retreat reworked and covered lateral moraine by proglacial deposits that were deeply eroded during the break of the terminal moraine dam of Lake Albany (~15 ka); possible interruption of deglaciation by climate change caused by the release of cold water from Lake Albany into the Atlantic and continued ice retreat recorded by glacioluvial, cross-beded deposits.

- The combination of methods used in this study proves to be an invaluable tool in glacial research in developed urbanized areas, such as the lower Hudson Valley. Specifically, LiDAR and geophysical surveys helped to map and interpret Jones Point terraces and stratigraphy with high spatial resolution. Future work will be planned to recognize and characterize more of these features using LiDAR data.

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Appendix A. Supplementary data

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References
