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When ice meets bedrock: variation in regional lithology as a control on the size variation of the Finger Lakes in Western New York

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Abstract The Finger Lakes are a set of eleven elongate lakes in Western New York that were carved by glaciers at the edge of the Laurentide Ice Sheet and are one of the most iconic glacial features in the world. Despite their common genesis, the lakes vary greatly in their length and depth. This study investigates the hypothesis that variation in the regional bedrock, rather than differences in ice flow patterns, may have been responsible for this size variation, with the longest and deepest lakes occupying an optimal location in easily erodible bedrock. To test this hypothesis, a simple one dimensional (1D) numerical model of glacier advance and erosion was used to find that lithologic changes alone can explain first-order Finger Lakes size differences using three lakes as case studies. Results also show that lithology can explain the observed bathymetric morphology of the lakes, with the deepest points of each lake occurring approximately two-thirds of the way down the trough from the northern shore. Regional lithology may thus provide an explanation for the dimensions and profiles of the Finger Lakes with likely implications for other glacially carved landscapes.

Non-technical summary The Finger Lakes are a set of eleven elongate lakes in Western New York that were carved by glaciers tens of thousands to millions of years ago. They are one of the most iconic glacial features in the world. Despite their common origin, the lakes vary greatly in their length and depth. This study investigates whether variation in underlying rock type, rather than differences in glacier ice flow patterns, may have been responsible for this size variation, with the longest and deepest lakes occupying an optimal location in easily erodible bedrock. To test this hypothesis, a simplified computer model of glacier erosion was used using three lakes as case studies. Results show that rock type changes alone can explain Finger Lakes size differences. Results also show that rock type can explain the observed shape of the lakes, with the deepest points of each lake occurring approximately two-thirds of the way down from the northern shore. Regional rock type may thus provide an explanation for the dimensions and profiles of the Finger Lakes with likely implications for other glacially carved landscapes.

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1 Introduction

The Finger Lakes region in Western New York is a paradigmatic example of a landscape sculpted by glaciers. Bounded by Lake Ontario to the north and the Allegheny Plateau to the south, the Finger Lakes region houses numerous distinctive glacial landforms left behind by the advances and retreats of the Laurentide Ice Sheet over the region during the last Ice Age (Clayton, 1972; Batchelor et al., 2019; Dalton et al., 2020, 2023), the latest of which occurred in the Late Wisconsin Glaciation (roughly 10–30 ka) and wiped away much of the evidence of earlier glaciations (Bloom, 2018). The Laurentide Ice Sheet consisted of many individual ice lobes that followed distinct flow routes (Margold et al., 2015). Of these, it was the southwestward trending Ontario Lobe that passed over the Finger Lakes region (Karig and Isacks, 2019; Bloom, 2018). Individual outlet glaciers stemming from the lobe filled in the various valleys, forming a complex pattern of flow (Clayton, 1972). The flow of the ice sheet over New York was influenced by the pre-existing topographic relief; for example, in the rugged terrain of intersecting low-standing valleys and upland plateaus, with reliefs commonly reaching 300 m established by pre-existing river drainage systems, the ice reached its peak thickness and flow speeds in the valleys and thinned out and slowed down atop the plateaus. Glacial denudation was thus preferentially concentrated in the valleys, with little significant modification of the plateaus (Clayton, 1972; Nair, 2018). At its maximum extent, the Laurentide Ice Sheet would push just beyond the southern border of New York into Pennsylvania (Karig and Isacks, 2019; Bloom, 2018).

The Finger Lakes, after which the region is named, are a roughly parallel set of eleven thin, elongate glacially-eroded lakes with a broader, more bulbous north shore and a thinner south shore (Fig. 1a; Mullins and Hinchey, 1989). The lakes are bounded on either side by steep bedrock walls (Fig. 1b), with reliefs between the deepest parts of the Cayuga and Seneca lakes

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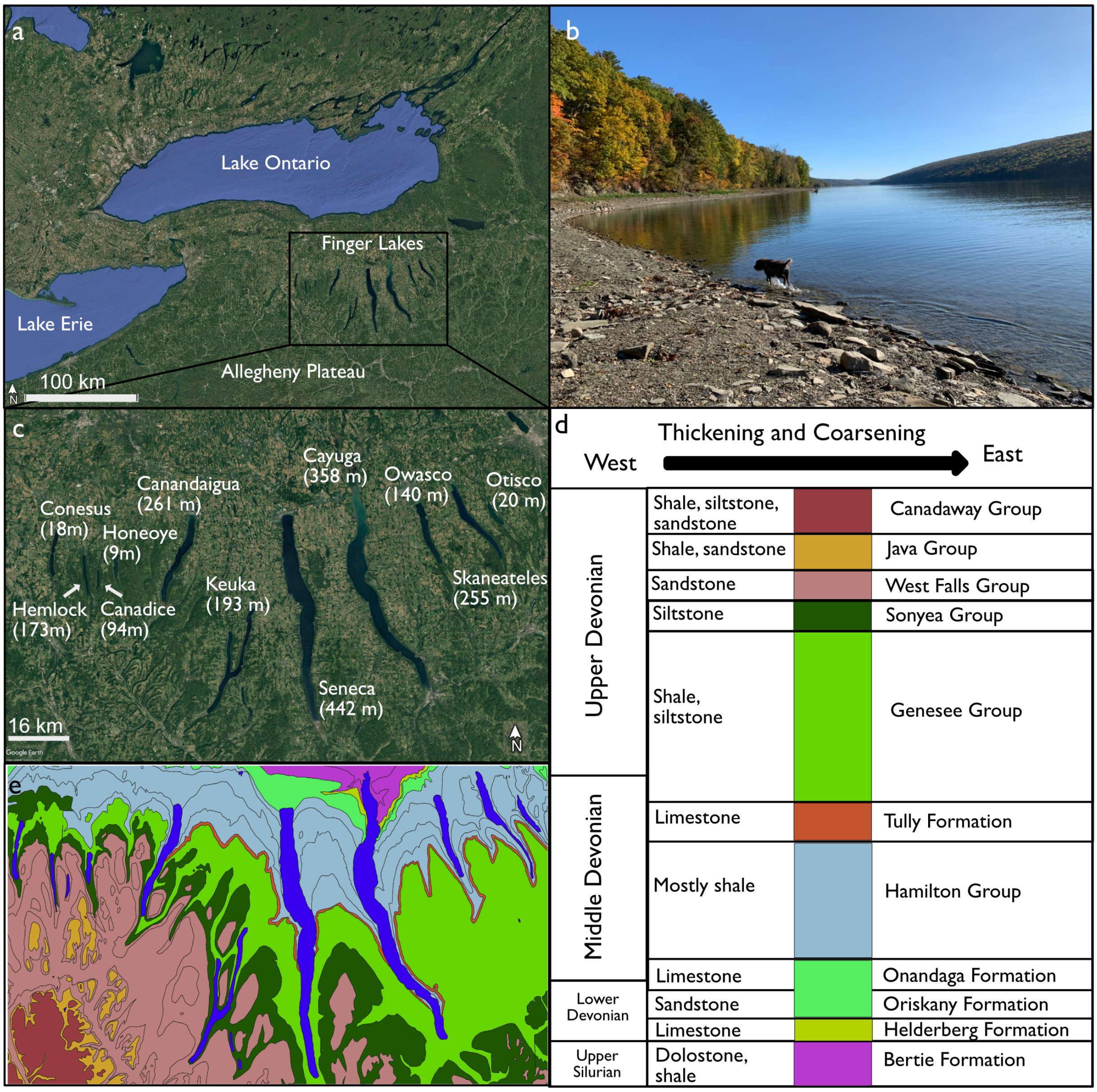

Figure 1 Morphology and geology of the Finger Lakes. **a**) Google Earth image of the Finger Lakes within the broader context of the region. **b**) Photograph of Hemlock Lake illustrating topographic setting. **c**) Google Earth image of each of the Finger Lakes with associated depth to bedrock (after Mullins et al., 1996). **d**) Detailed stratigraphic column of rock layers present within the Finger Lakes illustrating eastward coarsening of layers. Colors correspond to map in part e. **e**) Geologic map of the Finger Lakes (data from New York State Museum). See part d for legend.

trough and the highest bedrock ridges going over 900 m (Bloom, 2018). These steep basins extend 70 km southwards into the Allegheny Plateau, and the drainage divides at the southern ends of these basins that separate the St. Lawrence River Basin to the north and Susquehanna River Basin to the south have been glacially eroded to characteristically minimal reliefs, garnering them the name of through valleys (Tarr, 1894, 1905; Holmes, 1937; Muller, 1965; Muller and Prest, 1985; Muller and Cadwell, 1985; Cadwell and Muller, 2004; Bloom, 2018; Sookhan et al., 2021; Karig, 2022).

The south ends of the lakes are also dammed by the Valley Heads Moraine (VHM), a 280 km-long band of sediment deposited by glacial meltwater (Yang, 1992). The VHM, which has been dated to 17000 ka, was deposited during an intermittent glacial readvance onset by large-scale surge activity punctuating the general trend of the retreat of the ice sheet from the Finger Lakes region (Fullerton, 1986; Muller and Calkin, 1993; Ridge, 2003; Karig and Isacks, 2019; Karig, 2023). Retreat of the ice exposed the lakes to more drainage outlets to the north, which lowered lake levels until all that remained were the thin, elongate Finger Lakes (Cadwell and Muller, 2004; Bloom, 2018; Mullins et al., 1996; Kozlowski and Graham, 2014).

Previous work on the origin of landforms in the Finger Lakes region has focused principally on the role of pre-glacial river patterns and flow dynamics of the ice sheet (Fenneman, 1938; Flint, 1958; Monnett, 1924; Clayton, 1972; Ridky and Bindschadler, 1990; Cadwell and Muller, 2004; Bloom, 2018). Yet despite their common genesis, the stark size variation of the lakes has not yet been explained. The lakes range from the 5 km long Canadice Lake up to the 61 km long Cayuga and Seneca Lake (Fig. 1a; Mullins et al., 1996). The lake bottoms are filled with unconsolidated sediments that were deposited in waves by pressurized subglacial waters as the glaciers receded (Mullins and Hinchey, 1989), so that the bedrock elevation in Cayuga and Seneca Lakes extends as deep as 242 m and 306 m below sea level, respectively, making them far deeper than the other lakes (Fig. 1a; Mullins et al., 1996; Sookhan et al., 2021). Despite the size variation, the lakes share a common bathymetric pattern in that they all deepen southward and reFig. 1a;n to the southern shore (Mullins and Hinchey, 1989; Bloom, 2018). This study uses a 1D numerical model of glacial erosion in the presence of Finger Lakes lithology (represented by rock layers with differing thickness and erodibility) to examine how lithology might have affected the evolution and final shape of the lakes.

2 Geologic Setting

The Finger Lakes are cut through gently southward-dipping Devonian-age strata (Fig. 1b–c; Mullins and Hinchey, 1989). The oldest of these is the resistant Onondaga Limestone, which outcrops towards the northern shores of the lakes. Overlying this unit are two shale-rich groups, the Hamilton Group and Genesee Group, which sandwich the thin Tully Limestone. Towards the southern shores of the lakes and in the Allegheny Plateau, the silt- and sandstone-rich Sonyea Group and West Falls Group outcrop. The full details of the lithology of each of the strata are given in the Supplementary Materials.

Notably, the Devonian-age bedrock units thicken and coarsen, with the siliciclastic units becoming siltier and sandier, and likewise less erodible, going west to east (De Witt et al., 1993; Bloom, 2018). The Hamilton Group, for example, is 760 m thick in Ulster and Greene County towards the east but thins to just 75 m near Lake Ontario (De Witt et al., 1993). In the Genesee Group there is also the grading of the siltstone-rich Sherburne Flagstone and Ithaca Formations east of Seneca Lake (Fig. 1c De Witt and Colton, 1978). This lateral variation results from the depositional conditions of the bedrock layers. Namely, the bedrock derives from sediments that were shed from the Appalachian Mountains as they were uplifted by the Acadian Orogeny during the Middle Devonian Period (Williams and Hatcher, 1982; Ettensohn, 1985). These sediments accumulated in the Catskill Delta, which prograded westward over the epicontinental sea that covered Western New York, depositing a succession of thinner and finer sediment layers with the reduction in depositional energy (Ettensohn, 1985; Bloom, 2018). Additionally, at this time, a deep trough formed in the region corresponding to the Cayuga and Seneca Lake basins, allowing for a locally thicker accumulation of bedrock, especially shale, in this region (Bloom, 2018).

This study investigates the hypothesis that the largest lakes, Cayuga and Seneca Lake, fall within an opportune longitudinal band, since it is here that the bedrock, particularly the Hamilton and Genesee Groups, is both thick and fine-grained enough to allow substantial erosion before reaching the resistant Onondaga Limestone (Bloom, 2004, 2018). Where these bedrock layers are either too thin, as they are west of the two lakes, or too coarse and resistant to erosion, as they are east of the two lakes (De Witt and Colton, 1978; De Witt et al., 1993), the lakes are not as deep. While the exact geographic extent of this lithologic "sweet spot" is not well known, this study explores this hypothesis by implementing a 1D numerical glacier model in the presence of bedrock layers spanning the lithologic gradients from east to west representative of Canandaigua, Cayuga, and Skaneateles Lakes.

3 Methods

To explore the role of bedrock lithology in the formation of the Finger Lakes, this study uses a simple 1D finite difference model (using MATLAB) of glacier flow along a north-south-oriented line over a region with bedrock that mimics that of the Finger Lakes region. This study does not attempt to accurately simulate large scale ice sheet dynamics (e.g., Gowan et al., 2021) or accurately reconstruct paleo landscape evolution (e.g., Naylor et al., 2021); rather, the aim of this simplified "toy" model is to explore the possibility that lithologic differences alone can explain the first-order morphology and significant variation in size of glacially carved lakes, using the bedrock lithology of the Finger Lakes as a case study.

The simple numerical model follows many numerical flowline models for glacier flow and erosion (Harbor, 1992; Oerlemans, 1994; Jóhannesson, 1997; MacGregor et al., 2009; Anderson, 2014; Delaney et al., 2023). The model includes an initial topography and bedrock layers with distinct erodibility, accounts for mass balance by incorporating appropriate rules for glacier dynamics and ablation, and incorporates a simple erosion rule. While it likely took several glacial occupations to fully carve out the Finger Lakes (Bloom, 2018), for simplicity only a single occupation occurs in this model. This is accounted for by intentionally using erodibilities for the bedrock layers in the erosion rule that are much higher than estimates in real landscapes (e.g., Humphrey and Raymond, 1994), which is acceptable because the model aims to explore the effects of relative differences in erodibility between rock units, not absolute values of erodibility. This results in model times that are much shorter than the estimated actual time it took for the carving of the Finger Lakes. No attempt is made to account for the deposition of the lakefill sediments or the Valley Heads Moraine; as such, the final lake profiles represent the bedrock profiles in the absence of subsequent deposition.

This model explores several cases that include using bedrock layer thicknesses and erodibilities that correlate with Cayuga Lake, which occupies the proposed sweet spot of easily erodible rock, Canandaigua Lake, which lies west of Cayuga and Seneca Lakes and has thin, fine-grained bedrock layers, and Skaneateles Lake which lies east of Cayuga and Seneca Lakes and has thick, coarse-grained bedrock layers. Each part of the model is described in the following sections. Complete model scripts and a table of model parameters (Table S1) can be found in Supplementary Materials.

3.1 Spatial and Temporal Scale

The spatial domain includes an x-axis that represents horizontal distance along a north-south oriented line and a z-axis that represents elevation. The model runs over a horizontal distance of 120 km, roughly double the length of Cayuga and Seneca Lakes, and the grid spacing, dx, is set to 100 m. While the actual distance between the ice accumulation center in Hudson Bay and the Finger Lakes is around 1000 km, including it would lead to an unreasonably large spatial domain and model runtime, hence the selection of the smaller domain. The model runs for 4000 years, and, due to the nonlinear nature of ice flow, a short time step of $dt = 10^{-4}$ yr, or about $\frac{1}{25}$ of a day, was selected to ensure numerical stability.

3.2 Initial Topography

The initial topography starts at 1500 m, from which it exponentially decays to 600 m over a distance of 35 km (Fig. 2). Beyond this, the topography is linear and gently slopes down to 450 m at the end of the domain. The gently sloping region is meant to represent the actual Finger Lakes region, which has plateau elevations in the same range, whereas the exponential region is a fictitious region that acts as a horizontal contraction of the massive distance between the ice-accumulation center and the Finger Lakes. The selection of an elevation profile that slopes downward to the south tallies with the hypothesis that the pre-glacial rivers in the region flowed southward (Clayton, 1972; Bloom, 2018), while having the added benefit of improving model stability by preventing the glacier from encountering a back slope at the transition point from the exponential to the linear region.

3.3 Bedrock Layers

Bedrock layers are set in the model by plotting the line that represents the bottom of each bedrock layer (Fig. 2). Due to the gentle dip of the layers in the Finger Lakes region, this results in a series of lines with shallow, negative slopes. In determining the thicknesses and depths for each bedrock layer, for Cayuga Lake, a geologic cross-section of the lake based on seismic imagery was already available in Mullins et al. (1991). This cross-section was used to determine the elevation of the bottom of each of the bedrock layers at some distance along the length of the lake. For Canandaigua and Skaneateles Lake, geologic cross sections are available, though they do not distinguish or show all the bedrock layers that are needed for the initial settings of the model. Instead, KML (Keyhole Markup Language) files of bedrock geology data from New York State Museum (Fig. 1b) were used to determine the elevation of bedrock outcrops and the extent of the outcrop at the surface, which may be related to layer thickness due to the dip in the region being known. The collection of bedrock data was done along a north-south trending line in the vicinity of each lake, and the northernmost position at which the bedrock layer outcrops along that north-south line was recorded. This northernmost point is set to occur at $x = 45$km, in the linear topographic region of the model domain. While the transition from exponential to linear occurs at $x = 35$ km, an additional runout distance of 10 km is allowed in order to dampen the velocity buildup that the glacier has in the high-slope exponential region.

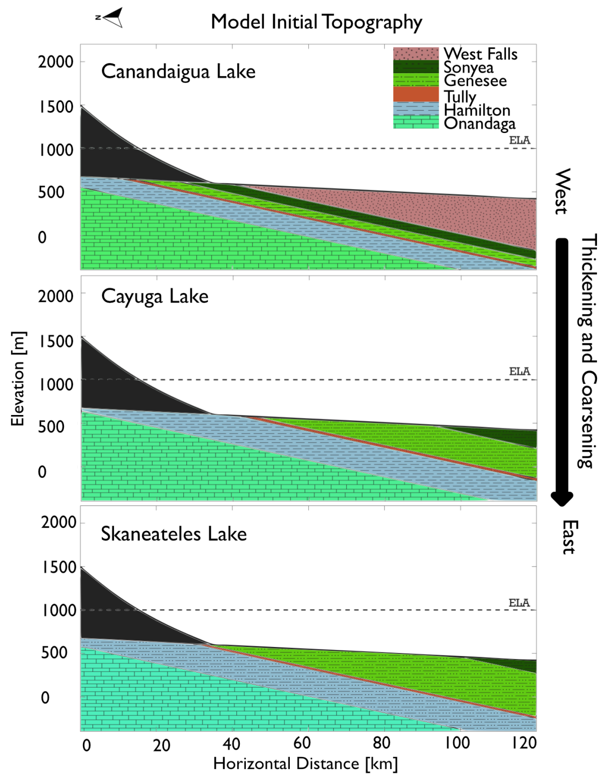

Figure 2 Initial topography, bedrock layers and C_2 values for each model case. Bedrock units thicken and coarsen (decreasing C_2) moving from west to east (top to bottom).

For all the models, the bottom of the Onondaga Limestone is not defined, and neither are the bottoms of the sub-Onondaga layers. The reason for this is that none of the real lakes in the study cut beyond the Onondaga Limestone (Mullins et al., 1991, 1996), and, as is to be shown in the model results in Section 3, none of the model lakes cut too deep into the Onondaga Limestone so as to warrant the need to assign another bedrock layer underneath the Onondaga (Figs. 4,5,6). The exact thicknesses and elevations of each of the bedrock layers that were used in the models are listed for each of the cases in Supplementary Materials (Tables S2-4).

3.4 Climatic Conditions

The climatic conditions of the model are encapsulated by the definition of an equilibrium line altitude (ELA), an elevation at which, locally, there is no net accumulation of ice. The ELA in the model is defined to be 1000 m based on cirque floor elevation data at the latitude of New York (Pelto, 1992). When the ice in a cell is above the ELA, there is net accumulation, while below there is net melting, with the magnitude being determined by the elevation differential between the ice and the ELA, such that:

Table 1 C_2 rock erodibility values used in this study. Lakes are listed left to right from west to east.

Stratigraphic Unit	Canandaigua C_2	Cayuga C_2	Skaneateles C_2
West Falls Group	0.45	N/A	N/A
Sonyea Group	0.90	0.90	0.50
Genesee Group	1.50	1.10	0.50
Tully Limestone	0.50	0.50	0.50
Hamilton Group	1.70	1.50	0.80
Onandaga Limestone	0.20	0.20	0.20
Exponential Region	1.00	1.00	1.00

$$b = \frac{db}{dz}\ z - \mathrm{ELA} \tag{1}$$

Here, b refers to the local mass balance in metres of snow per year and $\frac{db}{dz}$ is the rate of change in the snow balance with respect to elevation. For this model, $\frac{db}{dz}$ was set to 0.01 and b was capped at 2 m yr^{-1}. Sensitivity testing with model runs with various other ELA elevations may be found in Supplementary Materials; differences in ELA do not change the major findings of this study.

3.5 Glacier Dynamics

The motion of the glacier consists of two components: an internal deformation component and a sliding component. In regards to the first component, ice undergoes plastic deformation when on a slope under the action of its own weight. The basal shear stress $_b$, the component of the ice weight that is parallel to the slope it is on, is defined at the cell edges to be:

$$\tau_b = \rho_i g S H_{\mathrm{edge}} \tag{2}$$

Where $_i$ is ice density, g is gravitational acceleration, S is the slope of the ice surface, and H_{edge} is the height of the ice interpolated at the cell edges. This shear stress is related with the strain of the ice with Glen's Flow Law (Glen, 1958) to yield the average deformational velocity at the cell edges u_{def}, namely:

$$u_{def} = \frac{A}{5}(\rho_i\, g\, S)^3 H_{edge}^4 \tag{3}$$

Here, A is the flow-law parameter, which has been found empirically to vary with the temperature (van der Veen and Whillans, 1990). For this study, however, it is set to a constant value of $2.110^{-16}\, Pa^{-3}\, yr^{-1}$. The sliding component of the glacier velocity results from the meltwater that accumulates and flows at the base of the glacier. The sliding velocity uslide is related to the basal shear stress (Weertman, 1957) and is defined at the edges of the cells as:

$$u_{slide} = C_1 \tau_b^m \tag{4}$$

C_1 is a sliding constant that is set to $5.110^{-6} m^2 s^2 yr^{-1}$ in this study. For simplicity the exponent m is set to 1, following previous research on subglacial erosion (Delaney et al., 2023) that proposes a linear relationship between sliding velocity and basal shear stress. The ice discharge, Q, sums together the two velocity components, such that:

$$Q = (u_{def} H_{edge}) + (u_{slide} H_{edge}) \tag{5}$$

The boundary conditions set on the first and last cell edges prevent ice from flowing into the domain, but allow the ice to flow out of it. There is debate about whether ice movement over the Finger Lakes region was primarily driven by deformation or sliding. A conceptual model for the deposition of lakefill sediments seem suggest the presence of ice streams (Mullins and Hinchey, 1989), which move primarily by sliding induced by high-pressure subglacial waters, while an ice sheet model to predict the position of moraines along glacial flowlines in the Finger Lakes region suggests the presence of outlet glaciers, which move primarily by deformation, due to higher basal shear stresses more accurately reproducing the position of moraines seen in the region (Ridky and Bindschadler, 1990). Because the simple numerical formulation aims to explore relative differences in glacial erosion between areas of different lithology, differentiating between different models of ice flow is beyond the scope of this study.

3.6 Mass Balance

The mass balance combines together the local mass balance and discharge, such that the change in the height of the ice in each cell per year, $\frac{dH}{dt}$, is defined as:

$$\frac{dH}{dt} = b + \frac{dQ}{dx} \tag{6}$$

As an additional condition, the height of the ice is capped at 2000 m so as to more accurately reflect the height of the ice over the Finger Lakes region (Bloom, 2018). If the ice in any cell goes beyond this cap, the excess height above 2000 m is subtracted from the glacier height over its entire length.

3.7 Erosion Rule

Some studies suggest a nonlinear relationship between erosion rate and sliding velocity, especially in abrasion dominant systems (Herman et al., 2015). Due to uncertainty of the dominant processes that formed the Finger Lakes, the model used here does not distinguish between abrasion and quarrying processes but rather simply scales erosion (ε) in each cell linearly with the sliding velocity (MacGregor et al., 2009; Herman et al., 2021; Delaney et al., 2023), such that:

$$\dot{\varepsilon} = C_2 u_{slide} \quad (7)$$

The constant C_2 is a constant that represents the erodibility of the bedrock. While abrasion and quarrying are distinct physical processes with their own rates of denudation and dependencies (Bernard, 1979, 1996; Iverson, 1995; MacGregor et al., 2009; Anderson, 2014)—such as how reverse slopes obstruct quarrying but enhance abrasion—the constants in their mathematical definitions are no more well defined than C_2. For the sake of simplicity, these additional degrees of freedom are not included and a single constant erosion model is used.

3.8 Selection of C_2 Values

Because the values of glacial erosion constants for the Finger Lakes region are not well-defined, here they are selected based on whether they produce reasonable erosion patterns and on basic consideration of how erodible different lithologies are relative to others (Table 1).

For example, shale should necessarily have a higher C_2 than sandstone. The process of determining C_2 values for the bedrock layers for this study was mainly done by trial-and-error. An initial baseline for appropriate C_2 values for different lithologies was determined in the first model case, the Cayuga Lake Case (Fig. 2; Sec. 3.1), by seeing which C_2 values allowed for the formation of a lake with dimensions and bathymetric profile matching Cayuga Lake under the simple model conditions. For Cayuga, the Onondaga Limestone was assigned the lowest value at 0.2 due to its composition of resistant, hard limestone, which generally has low erodibility (Steinemann et al., 2020). The Tully Limestone was given a higher value of 0.5 due to it also containing calcareous shale layers (Baird et al., 2012). In regards to the three siliciclastic layers, the Hamilton Group was given the highest C_2 value at 1.5 due to it being the richest in shale. The Genesee Group in the vicinity of Cayuga Lake has a facies change in its shale layers to a composition rich in siltstone, silty shale, shaley siltstone, and fine-grained sandstone and also contains two siltstone-rich layers, the Sherburne Flagstone Member and Ithaca Member (De Witt and Colton, 1978). As a result, a lower value of 1.1 is given. The Sonyea Group is yet coarser than the Genesee Group and is thus given a value of 0.9. The exponential portion of the topography, not actually being part of the Finger Lakes region, is also given its own C_2 value of 1, as seen in Fig. 2. From there, these C_2 values were toggled up or down for the models for Canandaigua and Skaneateles Lakes to represent the facies change of coarsening and thickening from west to east (see Fig. 2 for C_2 values used).

Sensitivity testing of C_2 with additional model runs with various C_2 values may be found in Supplementary Materials. C_2 values were chosen not to be representative of realistic values, which are unknown, but to explore whether relative differences in lithology alone can explain the size variation and morphology of the Finger Lakes. However, the range in values used in this study can be quantitatively compared with an extensive database of natural rock erodibility measurements (Haag and Schoenbohm, 2025). In our study, the maximum difference in C_2 is just less than an order of magnitude, where the Hamilton Group shale is 8 times more erodible than the Onandaga Limestone. While there is significant variability in erodibility within each type of natural rock, the measured maximum difference between the most erodible shale and the least erodible limestone is a factor of > 400 (Haag and Schoenbohm, 2025). Thus the range of our chosen C_2 values is well within that of natural rock. Because they are calibrated based on model results and not entirely determined a priori, C_2 values are further discussed in the Results and Discussion sections.

4 Results

Overall, model results illustrate that differences in lithology alone are enough to explain observed morphology and relative differences in length and depth among Cayuga, Canandaigua, and Skaneateles Lakes in the face of identical glacial advances. Comparison with a control case in which all layers have equal erodibility shows that lithology is indeed needed to reproduce observed lake morphologies. Comparison of lake length and maximum depth eroded below sea level with the measured dimensions of each lake (Mullins et al., 1991) show general agreement. Additional model runs further exploring the role of bedrock thickness and erodibility are provided in Supplementary Material Section III.

4.1 Model Evolution

For all three cases, as the model runs (Fig. 3), the glacier first accumulates in the exponential region, taking on a lobate morphology. By 600 years, it has advanced to the transition point from the exponential to the linear region, at which point it comes into contact with the Hamilton Group. Due to its high erodibility, the glacier quickly burrows through this layer only to be stymied from further deepening by the Onondaga Limestone below. Since the erosion happens so quickly, the glacier makes a steep cut into the bedrock on the down-glacier end.

As the glacier continues to build and advance farther down the domain, it contacts the more resistant Sonyea and Genesee Groups. In the Cayuga and Skaneateles case, the reduction in erosion rates here due to the resistance of the layers and the more gradual topography causes the steep cut in the bedrock to become more gently sloping as it propagates forward (Fig. 3 and S1). This is because it takes more time for the bedrock to erode down at any point, and during that time, the glacier can advance and start eroding the bedrock farther down in the domain. The result is that the lowest elevation that the glacier erodes down to does not occur near the maximum advance of the glacier, as would be expected if the trend at $t = 600$ yr was to continue. Instead, the point of the lowest elevation becomes more separated in distance from the maximum advance of the glacier with each snapshot in time, giving a profile that is reminiscent of the Finger Lakes, which reach their max-

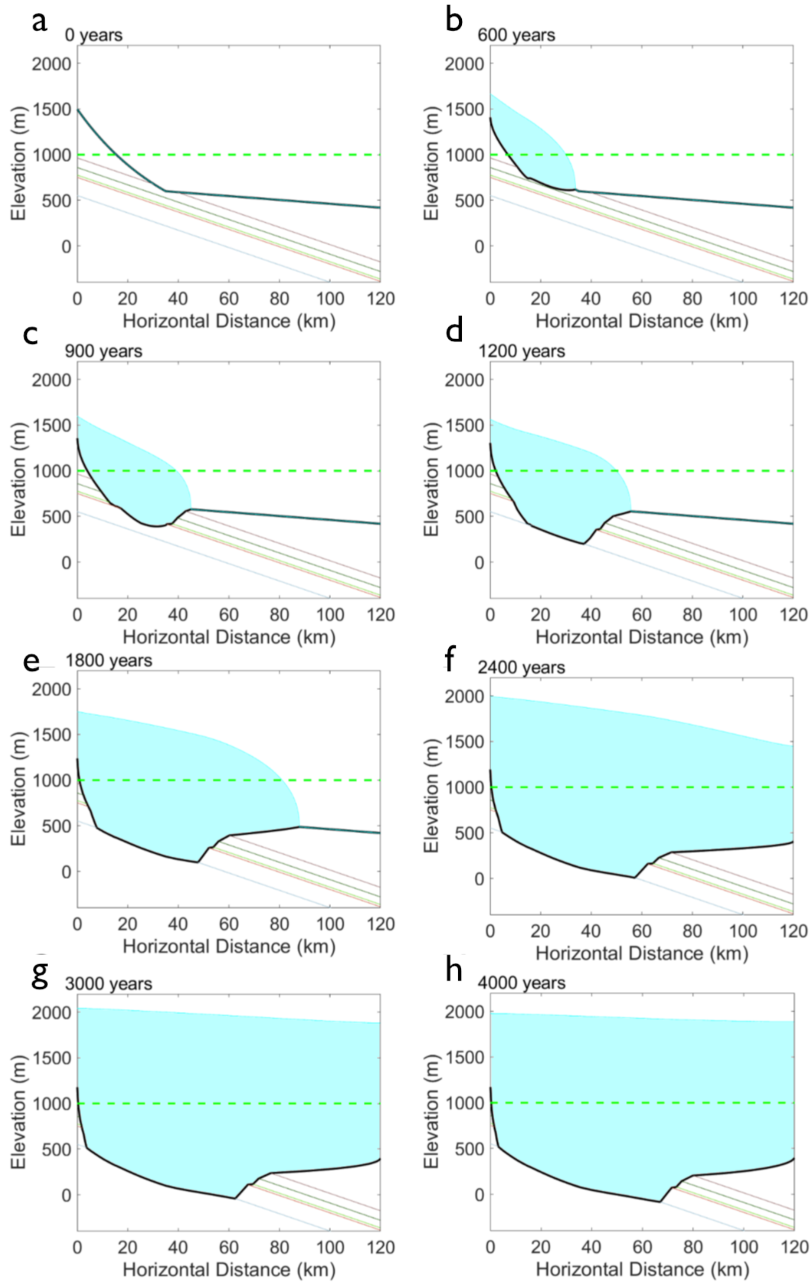

Figure 3 Time-lapse of the glacier growing over and eroding the bedrock at select time-intervals in the Cayuga Lake Case. The cyan line is the top of the glacier. For the bedrock layers, the blue line is the bottom of the Hamilton Group, the red line is the bottom of the Tully Limestone, the light green line is the bottom of the Genesee Group, and the dark green line is the bottom of the Sonyea Group.

imum depth two-thirds of the way along their length (Figs. 4,5,6). The Onondaga Limestone appears to be critical to this southward deepening bathymetry. Its resistance serves to cap erosion; its southward dip means that the glacier can cut deeper before hitting the cap the farther south it advances. In strong contrast, a control case with uniform erodibility ($C_2 = 1$) results in a southward shallowing bathymetry (Fig. 7). Together these results imply the importance of the resistance of the Onondaga Limestone for the overall bathymetric pattern of the lakes.

4.2 General Morphology Results

Model results show that differences in lithology alone can explain the first-order variation in lake morphology, with Cayuga Lake being the longest and deepest, while Canandaigua and Skaneateles Lakes are shorter and shallower (Figs. 4,5,6). Overall, the modeled lakes compare well with the measured dimensions of the real lakes derived from seismic reflection lines (Mullins et al., 1991). To determine the final dimensions of the model lakes, the assumption is made that $x = 45$ km represents the north shore of the lake since bedrock layers are set relative to this point. A horizontal line, representing the lake surface, is drawn from the bedrock surface at $x = 45$ km up until where it once again intersects the topography. The model Cayuga Lake is ~62 km long and has a base elevation of ~221 m below sea level at its deepest point, while the actual lake is ~61 km long and has a base elevation of ~242 m below sea level (Fig. 4b). Likewise, the model Skaneateles Lake is ~23.5 km long and has a base elevation of ~5 m above sea level, while the real life lake is ~24 km long and has a base elevation of ~8 m above sea level (Fig. 5b). Finally, the model Canandaigua Lake is ~25 km long and has a base elevation of 84 m below sea level, while the real life lake is ~25 km long and has a base elevation of ~50 meters below sea level. Note that the relevant depth of the real lakes for comparison is the total depth to bedrock at the bottom of the lakefill, not the water depth. However, while the lengths and depth below sea level generally agree with measured dimensions, the absolute depth of the modeled lakes is shallower than their real life counterparts (Figs. 4,5,6) due to increased erosion at the northern shore, which is discussed in the following section.

4.3 Interpretation of lake morphology results

While the modeled Cayuga Lake closely matched the length and depth of its real counterpart (Fig. 4), the absolute depth of modeled Canandaigua and Skaneateles Lakes is shallower than the real lakes (Figs. 5,6). This is due to increased erosion on the north side of the lakes in all model runs, which can be seen by comparing the elevation of the bedrock at the north shore of the lake in the models versus the real lakes (Figs. 4,5,6). This may be due the simplified glacial erosion model used, as well as the choice of C_2 values and the nuances of the real lithology; this and other caveats about model results are discussed below for Canandaigua and Skaneateles Lakes.

It should be noted that, in order for the model to accurately generate Skaneateles Lake so that the final dimensions are not too large, noticeably small values of C_2 needed to be used. Of course, C_2 should be lower for the Skaneateles Case than the Cayuga Case, though the change in C_2 going from Cayuga to Skaneateles is significantly greater than the change going from Canandaigua to Cayuga, indicating a more intense facies change (Table 1). However, the opposite would be the case in reality, as the facies of individual members that constitute each of the layers show only minor coarsening between Cayuga and Skaneateles Lakes (De Witt and Colton, 1978; De Witt et al., 1993), with the more abrupt facies transition from a predominantly shale-rich to siltstone-rich composition in the Genesee Group occurring along streams tributary to Seneca Lake (De Witt and Colton, 1978). What is more notable though is that the proportion of the overall bedrock layer that the members of the Genesee Group constitute does vary rather substantially, since, while the siltstone-rich Sherburne and Ithaca members thicken going from Cayuga Lake to Skaneateles Lake, the shale-rich Geneseo, Penn Yan, and Renwick members thin down (De Witt and Colton, 1978). This could give some justification to the C_2 of the Genesee Group as a whole being a fair bit lower for the Skaneateles Case than for the Cayuga Case. Even so, the large changes in C_2 may in part result from compensating for the initial topography in all three cases being the same, even though the preglacial valley depths prior to the glacier carving out the lakes were not necessarily exactly the same.

Results show that while the Canandaigua case shows a good match with the 25 km length of the actual lake, at a depth of 84.23 m below sea level, the model lake is 33 m too deep (Fig. 6). The lake trough that the modeled Canandaigua Lake fills is noticeably longer than the lake itself. As can be seen in the geologic cross section of Canandaigua Lake, lakefill extends past the southern shore of the lake (Fig. 6b, S2). The base of a substantial length of the Valley Head Moraine also falls under the water surface (Fig. 6b). Because subsequent deposition of the lakefill and VHM are not modelled, the model lake should thus be much longer than the real Canandaigua Lake. This problem stems from the erosion being too intense towards the northern shore, hence the bedrock elevation there for the model lake being much lower than in the real Canandaigua Lake, and too minimal towards the southern shore (Fig. 6b). Raising the C_2 of the West Falls Group to allow more erosion there has the unfortunate consequence of overdeepening the lake since it allows the deepest point of the lake to continue propagating southward, where it is able to reach lower elevations before coming into the proximity of the Onondaga Limestone. The size of Canandaigua Lake is limited more so by the presence of the high ridges of the VHM at its southern shore (Fig. 6b), but even still, the shallowing out of the base of the VHM does indicate the reduction in magnitude of glacial erosion latitudes far more north than in Cayuga Lake. The outcropping of the resistive West Falls Group must have limited glacial denudation so as to allow for a high elevation for the top of the VHM in the region, allowing the VHM to effectively dam Canandaigua Lake, and to prevent southward propagation of the deepest point in the lake. As in the model, the outcropping of resistive bedrock layers south of Canandaigua Lake must have been crucial to restricting its size.

5 Discussion

While these simple model runs cannot precisely simulate the formation of the Finger Lakes, the fact that the general pattern of bathymetry and length can be found by changing lithology alone under identical glacial conditions gives confidence that lithology may control real lake dimensions. Further, comparison to a control case with uniform erodibility shows that lithologic differences are needed to produce realistic lake shapes and dimensions; with uniform erodibility the deepest point occurs toward the upglacier end of the lake. It is also interesting that a model of a single glacial advance produces a lake basin that resembles real observations without modeling any moraine or lakefill deposition. This indicates that simple interactions between ice flow, erosion, and lithology can produce glacial lakes without the effect of moraine damming, though moraine damming remains an important control on lake morphology in many cases, as evidenced by catastrophic flooding when moraine dams are compromised (e.g., Westoby et al., 2014).

While results clearly show that lithology alone can alter lake morphology in a simple glacial erosion model, such simple models should not be used to predict spe-

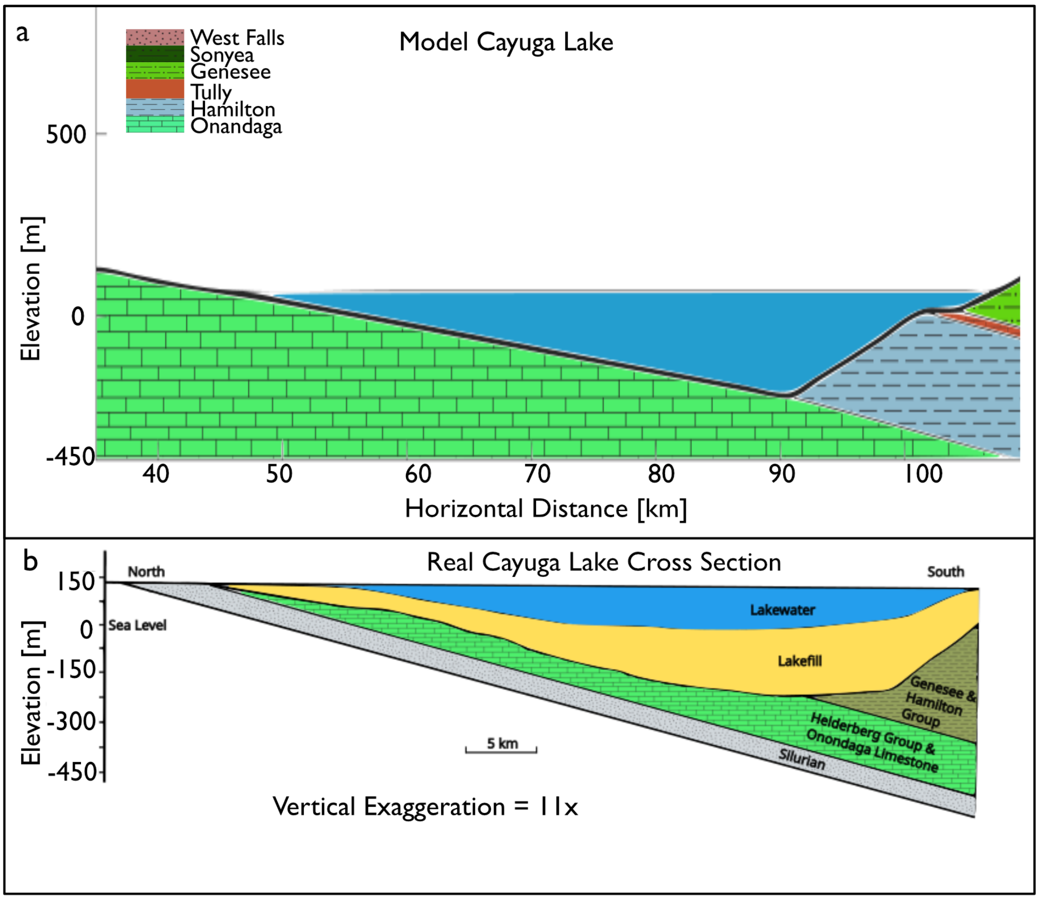

Figure 4 **a**) Model results for the lake that would fill the final trough in the Cayuga Lake Case. **b**) The geologic cross section and bathymetric profile of the real Cayuga Lake (adapted from Mullins et al., 1991).

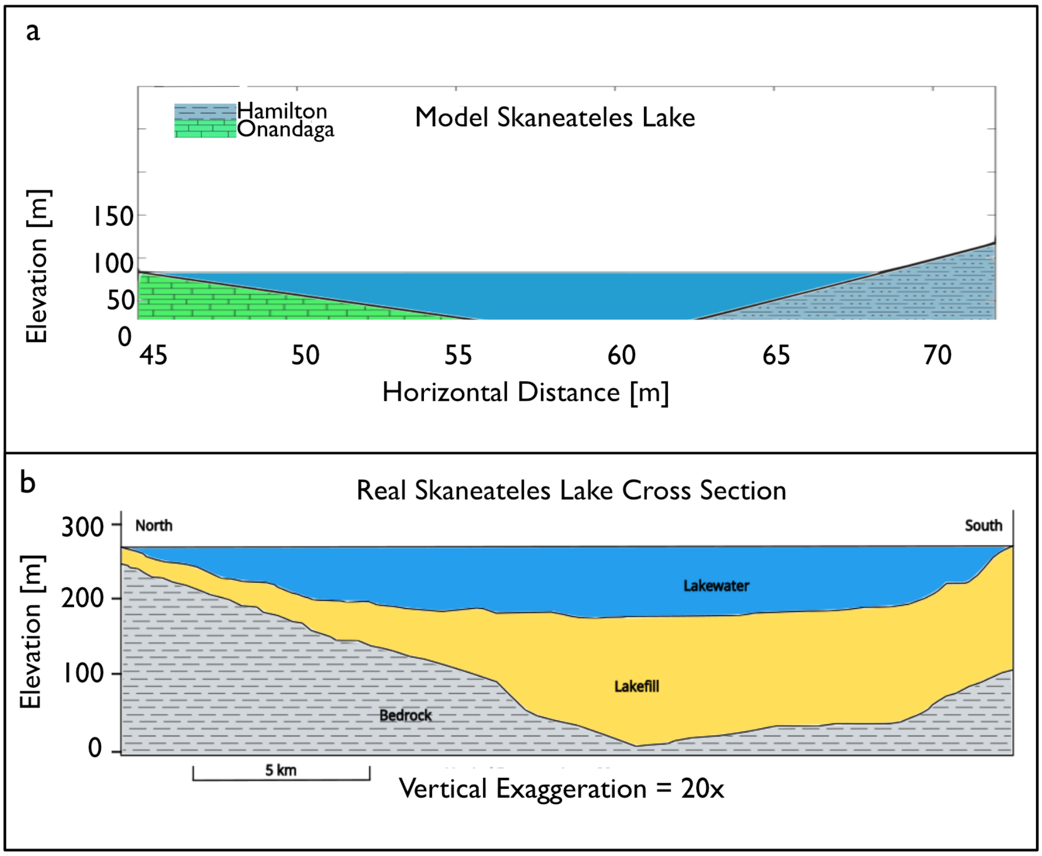

Figure 5 **a**) Model results for the lake that would fill the final trough in the Skaneateles Lake case. **b**) The geologic cross section and bathymetric profile of the real Skaneateles Lake (adapted from Mullins et al., 1991).

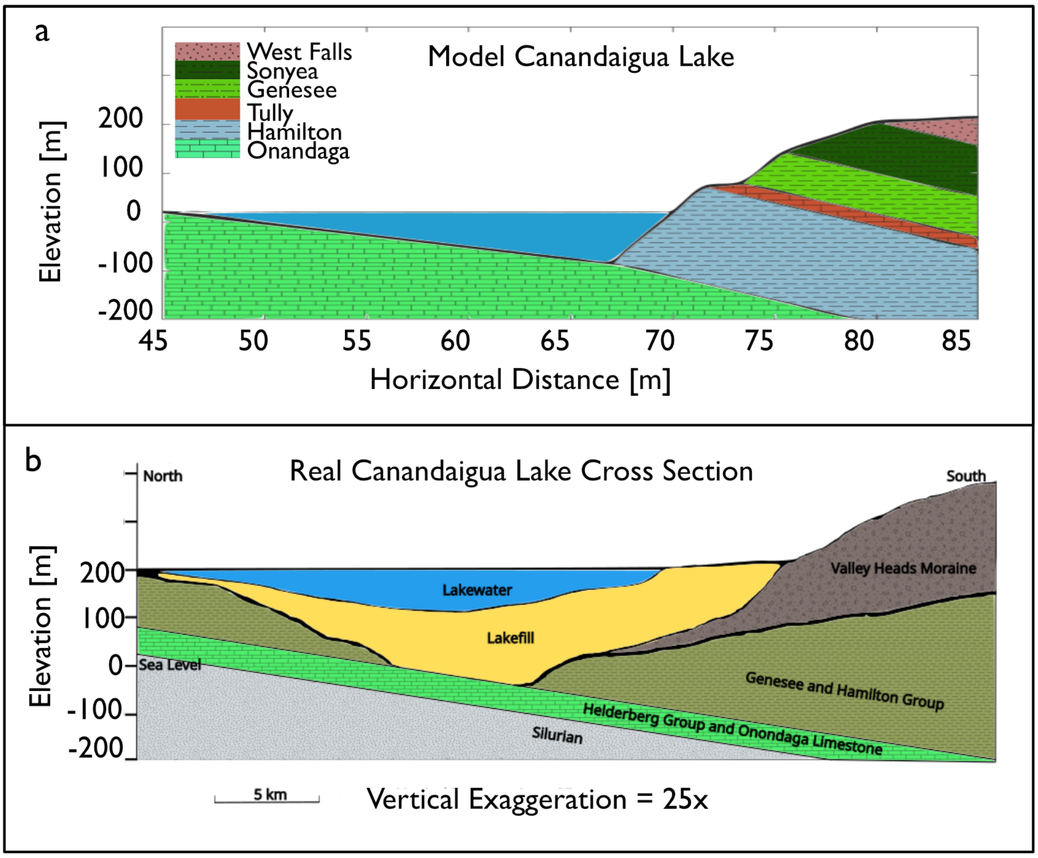

Figure 6 **a**) Model results for the lake that would fill the final trough in the Canandaigua Lake. **b**) The geologic cross section and bathymetric profile of the real Canandaigua Lake (adapted from Mullins et al., 1991).

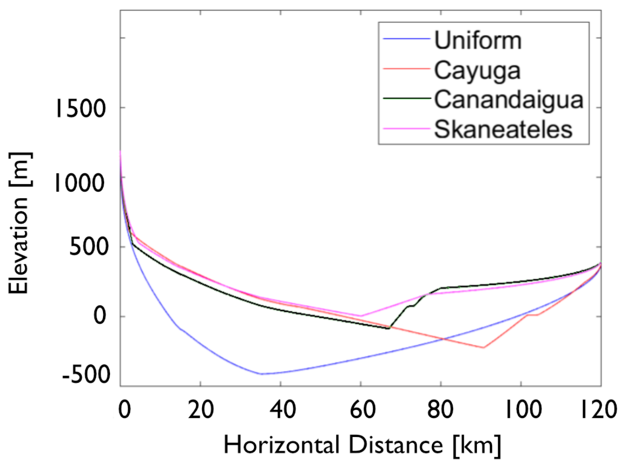

Figure 7 Comparison of final bedrock elevation for the three lake model runs and a control run with uniform erodibility

cific erosional conditions or simulate realistic glacial history in the Finger Lakes region. Recent advanced simulations of Laurentide ice sheet flow (Batchelor et al., 2019) along with improved dating methods (Dalton et al., 2020) show the path forward for high fidelity reconstructions of erosional history. Future simulations of glacial erosion in this region could combine advanced ice flow models with lithology dependent erosion rules that acknowledge the role of rock type.

The present study adds another dimension of analysis to the geomorphology of the Finger Lakes region by focusing on lithology instead of postglacial rebound (Miller and Douglas, 2006) or the flow patterns of the glaciers and pre-glacial rivers. While lithology may explain otherwise enigmatic features like the size variation and bathymetry of the Finger Lakes, it may also complement existing explanations for other features in the region that are based more on the flow dynamics of the glacier. For example, the lack of erosion south in the Allegheny Plateau has been explained as resulting from the thinning and slowing down of the ice sheet over the high plateau (Clayton, 1972). However, this may also be explained by the sandstone composition of the Allegheny Plateau, as it was seen that in the Canandaigua case that the West Falls Group experienced little denudation, despite the glacier not thinning over it, due to its low erodibility. On the Allegheny Plateau, the bedrock is primarily highly resistant sandstone that may act similarly to the highly resistant limestone in this study. In this way, the consideration of lithology can readily and harmoniously enter into a number of pre-existing conversations in Finger Lakes and glacial geomorphology.

These findings from a simple, generalized 1D model of glacial erosion suggest that lithology may exert a strong control on glacial erosion and postglacial landscapes around the world, beyond the Finger Lakes. In addition to the main effect of changing erodbility, lithology may influence glacial evolution in various ways. Lithology has been found to impact the distribution of rock glaciers in the western US (Johnson et al., 2007) by controlling clast size—which determines the insulating properties of the debris layer—and permeability of the bedrock, which controls the ability for surface water to persist. Resistant lithologies have also been observed to control glacier size in China due to orographic precipitation effects near high peaks (Schoenbohm et al., 2014). However, very few studies have explicitly explored the role of lithology in glacial evolution; this study suggests that future work in this area is warranted, with implications for our understanding of glacial erosion processes and rates, postglacial geomorphology, and lake geomorphology specifically.

6 Conclusions

The key findings of the study may be stated as fivefold:

1. The eastward thickening and coarsening of stratigraphic layers in the Finger Lakes region can account for their size variation. Lakes west of Cayuga and Seneca Lake (though not modeled in this study, Seneca has similar lithology and dimensions to Cayuga), such as Canandaigua Lake, are limited in their growth due to the thinness of the highly erodible, shale-rich bedrock, while lakes east of Cayuga and Seneca Lake, such as Skaneateles Lake, are limited in their growth due to the coarseness of the bedrock layers. The central longitudinal position of Cayuga and Seneca Lake in the Finger Lakes region provides an optimal middle ground in which the bedrock layers are both relatively thick and fine-grained (hence more erodible).
2. The Onondaga Limestone acts as a resistive underpinning that prevents much glacial erosion into it. The southward dip of the Onondaga Limestone governs the southward deepening of the lakes.
3. The deepest point in the lake occurs around two-thirds of the way down the lake from the northern shore because, as the glacier encounters the more resistant, younger bedrock layers that outcrop towards the south, the rate of erosion declines, allowing the lake to lengthen faster than it deepens.
4. The West Falls Group capped the southward lengthening of Canandaigua Lake and, moreover, limited the rate of erosion such that it dampened the enhancement in erosion potential in the region that was furnished by the finer facies of the underlying layers.
5. Glaciers may form a lake due simply to the spatial pattern of erodibility governed by the lithology. Deposition of a moraine to dam the lake is not required, though certainly aids in preservation of the lakes.

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Data and Codes Availability

All data and code used in this study are available in supplementary materials online. The supplementary movie can be found at https://doi.org/10.6084/m9.figshare.30044440.v1

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The authors declare that no artificial intelligence tool was used to assist in or generate any part of the contents in this manuscript.

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