UTILIZING LANDSLIDES IN LAKE CHAMPLAIN AS PALEOSEISMIC AND PALEOHAZARD INDICATORS

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ABSTRACT

Many landslides have been identified in lake sediments via Multibeam and CHIRP (compressed high intensity radar pulse) seismic profile imagery. Previous studies have shown that these landslides are coeval and were most likely triggered by a large earthquake ~4500 – 5500 years ago. These landslides also correlate with terrestrial landslides in Ottawa, for which a M 6.4 or greater earthquake trigger has been determined. Older landslides have also occurred, but no further investigation has been done.

This study focuses on a series of overlapping landslide deposits in an area between the Bouquet River Delta and Essex, NY in the main section of Lake Champlain, where nearly the entire slope has failed, with the exception of a few locations where blocks of sediment remain intact. However, individual scarp faces and debris fans are much better defined on the west side of the study area than the east, and this investigation with therefore focus on the western region. Core studies show that sedimentation rates are much higher on the west side of the lake than the east side because of sediment flux from the Bouquet River. Using sedimentation rate range of 0.13-0.17 cm/yr and the thickness of sediment accumulation above slumped material in seismic imagery, approximate failure ages have been calculated for each of the identified failed regions on the west side of the study area.

The northernmost failure occurred about 950-1200 years ago, and was the first mass wasting event of this age to be recorded on Lake Champlain. All three of the other regions experienced slope failure about 4500-5200 years ago, and these failures are also coeval with previously studied landslides on Lake Champlain.

In the nearby Western Quebec Seismic Zone (WQSZ), for example, clusters of terrestrial landslides occurred about 1 ka and 5 ka, and they are associated with a seismic event around the date of failure (Brooks, 2013). The landslides observed on Lake Champlain occurred along the same sediment interface, and were likely triggered by these same earthquakes.
To my parents, for believing in me no matter what.
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1. Introduction

Recent advances in remote sensing technologies such as Multibeam bathymetric mapping and seismic reflection profilers have made it possible to conduct in-depth studies of bottom and sub-bottom features in both marine and lacustrine environments (e.g. Chase and Hunt, 1972; Vogt and Smoot, 1984; Savini et al., 2011). Landslides, which are one such feature, have long been studied on land, but have much more recently become the focus of detailed subaqueous investigations (e.g. Schnellman et al., 2002; Gibert et al., 2005; Strasser et al., 2013). Such studies have important social implications because like their terrestrial counterparts, subaqueous landslides are a significant natural hazard. Throughout human history, surficial landslides have been responsible for significant infrastructure damage as well as the loss of human life. Recent research suggests that subaqueous landslides also present the potential for the additional hazard of landslide-generated tsunamis (Ward and Day, 2005).

New instrumentation on Middlebury College’s research vessel, the R/V David Folger, which was acquired in 2012, has allowed for the identification of several large underwater landslides beneath Lake Champlain that were previously unknown. Investigation of these landslides has identified several coeval failures that occurred around 5000 years ago, helping to shed light on the trigger mechanisms of landslides in the area (Ghosh, 2012; Rosales-Underbrink, 2015). However, little is known about paleohazards that may have been associated with these events, and the understanding of these hazards is essential in order to anticipate the potential effects of and plan for possible future slope failure.
2. Landslides: Background and Prior Research

2.1 Landslide Mechanics

A landslide is defined as a gravity-driven, down-slope movement of slopeforming material that occurs as the result of failure along one or multiple planes (Hampton et al., 1996). A slump is a type of landslide that is characterized by the downslope movement of an at least partially intact block of sediment, where original stratification may be compromised in places but generally remains intact (Sturm, 1971). Two diagnostic features of slumps are a rupture surface at the top of the slope and a displaced mass of sediment down below (Fig. 1) (Hampton et al., 1996). The rupture surface is the plane along which movement originated and is usually a concave-up rounded surface, while the displaced mass consists of the sediment that has been dislodged and has moved downslope (Hampton et al., 1996). A debris flow, on the other hand, is a type of slope failure where the original stratification of the displaced mass is not maintained (Felix and McCaffrey, 2005).

Figure 1. Schematic diagram of a typical slump structure, including a curved, concave-up rupture surface, a steep main scarp, a displaced sediment mass and a toe. From Ghosh, 2012, modified from Hampton et al., 1996.
2.2 Differences Between Terrestrial and Subaqueous Landslides

While there are many similarities between the structures of terrestrial and subaqueous landslides, a few key differences exist. First, terrestrial landslides typically occur on steep slopes ($\leq 10^\circ$), and while subaqueous landslides can occur on slopes of similar steepness, they can also occur on very shallow slopes with inclinations as low as $<1^\circ$ (Ward and Day, 2002). Examples include an earthquake-induced liquefaction flow on the Klamath River Delta off the shore of northern California which occurred on a slope of $0.25^\circ$, as well as a slide on the continental shelf in the Gulf of Alaska that occurred on a slope of $<0.5^\circ$ (Hampton et al., 1996).

Subaqueous landslides can also displace much larger quantities of material than terrestrial slides. For example, the largest known terrestrial landslide, which occurred on the slope of Mount Shasta in the southern Cascades Range in California, displaced approximately 26 km$^3$ of material, while the largest known submarine landslide displaced over 20,000 km$^3$ of sediment off the coast of South Africa (Hampton et al., 1996). This difference in size exists despite the fact that terrestrial landslides are much easier to identify and have been studied for a much longer period of time than subaqueous landslides, and it is therefore highly likely that even larger subaqueous slides have occurred but remain unknown to the scientific community.

2.3 Causes of Slope Failure

Factor of Safety (FS) is a dimensionless measure that relates resisting forces to driving forces (shear forces) in order to determine the likelihood of slope failure. It can be calculated using the following equation:
In both subaerial and subaqueous environments, a FS value greater than 1 indicates that the resisting forces exceed the shear forces acting on the slope, and it is therefore stable and unlikely to fail. A value of less than 1 indicates that the shear forces exceed the resisting forces and it is therefore likely that failure will occur.

This stress imbalance is often caused by internal deformation that results in either a decrease in resisting forces or an increase in shear forces. A decrease in resisting forces can be caused by hydraulic uplift due to pore fluid expansion (Sturm, 1971); the exolution of gas hydrate, which is typically associated with a decrease in water level (Felix and McCaffrey, 2005); or by preexisting weakness caused by the composition of the sediment (Locat and Lee, 2005). An increase in shear forces can be caused by processes such as rapid sedimentation, which causes the oversteepening of a slope and leads to failure when the inclination of the slope exceeds the angle of repose.

A trigger event is often necessary to cause this imbalance in resisting forces and shear forces. Triggers are typically stochastic in nature, where stochastic events include a wide variety of processes, such as episodic or seasonal precipitation events, groundwater seepage, volcanic activity, waves, tidal forces, and seismic activity (Locat and Lee, 2005). This investigation focuses on sediment composition and seismic activity, which have previously been determined to be the cause of many landslides in both Lake Champlain and the greater region (Ghosh, 2012; Rosales-Underbrink, 2015).
Earthquakes are a common trigger mechanism because the magnitude of the shear stress that they exert is quite large, and because their shaking triggers the movement of the pore water as well as the sediment, which causes instability (Locat and Lee, 2005). The presence of multiple slumps, debris flows, and turbidite sequences of similar ages within a region is indicative of a seismic trigger mechanism because it shows that the region as a whole was subjected to a specific event at the same time (Schnellmann et al., 2002).

2.4 Landslide Hazards

The study of surficial landslides has important social implications because of the dangers that they present to humans. The two main hazard concerns are loss of human life and infrastructure damage. In the 2015 landslide in El Cambray, Guatemala, which was triggered by an intense precipitation event, over 271 people were killed (Mullen and Ramos, 2015). In a landslide in Notre Dame de la Salette near Buckingham, Canada in the early 1900’s, 34 people were killed and nearly the entire town was destroyed (Duffy and Spears, 2008).

Evidence also suggests that subaqueous landslides pose an additional threat to human life by the potential for creating tsunamis. A tsunami is a surface wave generated by sudden displacement within or at the surface of a body of water, typically due to an earthquake, submarine landslide, surface landslide, asteroid impact, or a combination of these causes (Harbitz et al., 2006). Tsunamis are high-risk hazards and factors that must be considered in order to estimate their potential for damage include the spatial distribution and height of the wave (Harbitz et al., 2006). In the case of subaqueous
landslide-triggered tsunamis, these variables are influenced by the velocity and volume of
the landslide (Ward and Day, 2002; Harbitz et al., 2006). Tsunamis are difficult to plan
for on a long-term timescale and often strike with little warning and devastating
consequences. For example, a tsunami caused by a subaqueous landslide off the coast of
Papua New Guinea caused infrastructure damage and human casualties as far away as
Hawaii (Ward and Day, 2002).

Relevant to this study is past evidence of lake tsunamis associated with
seismically triggered lacustrine landslides. Historical records document a M 6.2
earthquake near Lake Lucerne in central Switzerland in 1601 AD (Schnellmann, et al.,
2002). C$^{14}$ dates on slump deposits within the lake basin indicate coeval subaqueous
slope failure around this same time, and suggest that these slumps were the result of a
seismic trigger (Schnellmann et al., 2002). In the same year, damage due to large waves
(up to 4 m) was also recorded in a small town on the shore of Lake Lucerne, and it can
therefore be inferred that the observed slumps triggered a lake tsunami when they failed
(Schnellmann et al., 2002). This study provides evidence that it is plausible for tsunamis
to have been generated by the seismically triggered mass wasting events that have been
identified in Lake Champlain, and also suggests that it could be possible for similar
events to occur in the future.

3. Regional Background and Prior Research

3.1 Lake Champlain

With a total lake surface area of 1,132 km$^2$ and a drainage basin area of about
21,326 km$^2$, Lake Champlain is the sixth largest natural lake in the United States (Manley
and Manley, 2009). It is located in the north-south oriented Champlain Valley, along the border between New York and Vermont and extends from Whitehall, New York all the way into the province of Quebec in Canada (Fig. 2). It is approximately 193 km long and has a mean depth of about 20 m, but is generally shallower towards the south and deepens gradually to a maximum of about 120 m near Split Rock Point (Freeman-Lynde et al., 1979). The lake drains to the north via the Richelieu River, which feeds into the St. Lawrence River (Freeman-Lynde et al., 1979). Lake Champlain can be subdivided into three sections: the Main Lake, the South Lake and the Restricted Arm (Fig. 2) (Manley and Manley, 2009). The Main Lake extends from the Crown Point Bridge in the south up to the northern extent of the lake in Canada, where it feeds into the Richelieu River (Manley and Manley, 2009). The Main Lake is the widest section of the lake, with a maximum width of 16 km near Burlington, Vermont and a typical width between 3 km and 5 km (Freeman-Lynde et al., 1979). The South Lake resembles a river in shape and is the narrowest region of the lake, with an average width of about 1 km (Freeman-Lynde et al., 1979). It extends from the Crown Point Bridge down to the southern point of the lake in Whitehall, New York (Manley and Manley, 2009). The Restricted Arm comprises the northeastern section of the lake and as its name suggests, water flow to and from this region of the lake is limited (Manley and Manley, 2009).
Figure 2. Bathymetric map of Lake Champlain with two different insets of the study area with an inset multibeam image of the study area overlain by the tracks of collected seismic lines. Black arrow indicates flow out of the lake and into the Richelieu River. Map modified from Manley et al., 2011.
3.2 Regional Sedimentation History

Three distinct depositional periods mark the history of the lake, each of which can be identified by distinct reflection packets in seismic profiles (Chase and Hunt, 1972; Cronin et al., 2008). In most places, these sediment packages overlie Paleozoic bedrock, but in some locations the bedrock is overlain by glacial till (Freeman-Lynde et al., 1980). These three depositional periods span the past 14 ka and each represents a distinct environment of deposition that corresponds with a different phase of the regional postglacial history (Fig. 3).

Figure 3. Maximum extents of Lake Vermont, Champlain Sea and Lake Champlain. From Ghosh, 2012, modified from LCBP, 2004.

3.2.1 Lake Vermont

Lake Vermont was part of a series of proglacial lakes that formed at the ice front of the Laurentide Ice Sheet as it retreated northward through the Champlain Valley towards the end of the Wisconsin glaciation, about 14 ka (Chase and Hunt, 1972; Cronin
et al., 2008). During the maximum extent of glaciation, ice sheets covered northeastern North America from northern Canada all the way down to Long Island. Towards the end of glaciation, as the ice sheet retreated northwards and exposed the Champlain Valley, meltwater accumulated in the recently deglaciated basin and formed a series of proglacial lakes that were known by a variety of different names as the extent of the area they covered shifted over time. Lake Vermont is the name given to two of these stages – the Lake Coveville stage and the Fort Ann stage. Sedimentation during this period occurred at a relatively rapid rate of ~4-8 cm/yr and consisted largely of the settling of glacial sediments that were transported into Lake Vermont via meltwater runoff (Freeman-Lynde et al., 1980; Cronin et al., 2008).

Seismic profiles of Lake Vermont sediments are dark and acoustically nondescript, with a sudden and well-defined transition to the overlying acoustic unit (Fig. 4) (Cronin et al., 2008). The Lake Vermont sediment displays significant variation in thickness, ranging from very thin or absent above bedrock points and near the shore, to more than 126 m thick near the deepest part of the lake just north of Split Rock Point (Chase and Hunt, 1972). The sediment itself is very dense, nonfossiliferous clay with alternating layers of dark brown and dark gray (Chase and Hunt, 1972).
3.2.2 Champlain Sea

The Champlain Sea was a marine estuary environment that existed between ~13-9 ka, when the Laurentide Ice Sheet had retreated to the north and the Atlantic Ocean inundated the isostatically depressed Champlain and St. Lawrence Valleys (Chapman, 1937; Cronin et al., 2008). Paleobeaches in the region indicate that relative sea level during this period was ~120-160 m higher than modern sea level (Fig. 3) (Chapman, 1937).

Champlain Sea sediments were deposited directly above Lake Vermont sediments, at a rate of approximately 1 cm/yr (Cronin et al., 2008). The transition between Lake Vermont and Champlain Sea sediment is easily recognizable in seismic profiles because the density difference between sediment units results in a clear acoustic reflector (Fig. 4) (Cronin et al., 2008). The characteristic seismic profiles of Champlain
Sea sediments are nearly transparent during the early phases of the Champlain Sea and become acoustically laminated towards the top of the sediment package (Fig. 4) (Cronin et al., 2008). The sediment itself is dark gray clay with black mottling (Ghosh, 2012). It has a higher moisture content and a lower organic content than more recently deposited sediments, and lacks diatoms, plant fibers and copepods, but contains an abundance of marine foraminifera, which is evidence of deposition in a brackish environment (Chase and Hunt, 1972). Physical property measurements show that the magnetic susceptibility is higher, density is greater and porosity is lower than more recently accumulated sediments (Cronin et al., 2008; Dawson, 2008).

3.2.3 Lake Champlain

After the complete retreat of the Laurentide Ice Sheet, the region isostatically rebounded and the confluence with the Atlantic Ocean was eventually cut off, forcing a transition back to a fresh water environment. The resulting body of water is known as Lake Champlain. In seismic profile, these transition period sediments are acoustically laminated and likely started to accumulate around the time of the first appearance of non-marine protozoans, which was about 9.7 ka (Cronin et al., 2008). The transition from Champlain Sea to Lake Champlain sedimentation is gradual in the deeper parts of the lake and abrupt along the shore (Cronin et al., 2008).

Seismic profiles of Lake Champlain sediments are acoustically laminated, but can be distinguished from the underlying laminated Champlain Sea sediments because they have a lower reflectivity (Fig. 4) (Cronin et al., 2008). Average sedimentation rate ranges from about 0.05 cm/yr to 0.15 cm/yr (Hunt, 1976; Cronin et al., 2008) and the sediment
itself exhibits substantial variability in composition, but is typically a diatomaceous, dark gray to grayish brown organic mud that is poorly sorted with grain size ranging from clay to gravel (Chase and Hunt, 1972).

### 3.3 Regional Landslide History

#### 3.3.1 Landslides beneath Lake Champlain

On Lake Champlain, many slumps and debris flows have been identified via Multibeam and CHIRP (compressed high intensity radar pulse) seismic profile imagery, and a total of seven have been studied in detail. The Diamond Island slump is located just south of Split Rock Point in the Main Lake near Ferrisburgh, VT and displaced about 100,000 m$^3$ of sediment down a 16 m near-vertical scarp (Manley and Manley, 2009). It is associated with a characteristic slump scar and downslope rotated block and the displaced material remained almost fully intact. Failure occurred at a transition from the outer portion of the shelf region to the slope region. The Whallon Bay slump, located in the Main Lake just north and west of Split Rock Point, displaced a total sediment volume of 4.0x10$^6$ m$^3$. Like the Diamond Island slump, it occurred at a transition from shallow to deeper basin depth and has a characteristic downslope rotated block, but has no visible detachment scarp and the displaced material has broken into three small, rotated blocks that lie at the toe of the slump.

The Folger Trough slump is located near Basin Harbor in the southern part of the Main Lake and has a small, ~5 m scarp with displaced material below that is disturbed, folded and acoustically transparent (Manley and Manley, 2009). Failure occurred along the interface between Champlain Sea and Lake Champlain sediments.
More recent studies have investigated the Four Brothers slump, which is located southwest of the Four Brothers Islands in the Main Lake and has a scarp of ~5 m, with a displaced body of material that is about 5.5 x 10^5 m^3 (Fig. 5) (Ghosh, 2012). Three large debris flows have also been identified near Quaker Point, northeast of the Four Brothers slump in the Main Lake. In each of these cases, the failure plane is located at the main interface between Lake Champlain and Champlain Sea sediments (Manley and Manley, 2009; Ghosh, 2012; Rosales-Underbrink, 2015; Manley et al., 2015).

Detailed analysis on the Four Brothers’ Slump and Quaker Point debris flows by Middlebury thesis students Jaya Ghosh (2012) and Piper Rosales-Underbrink (2015) suggest that these landslides were triggered by a large earthquake ~4500 – 5500 years ago. Evidence for a seismic trigger event comes from the dating of these landslides around ~5000 BP. These coeval landslides occur beneath water depths of 65-338 ft, thus below wave base, and the lack of shoreline geometry at these depths combined with the low local sedimentation rates (~0.06-0.07 cm/yr) suggest that slope failure is likely not due to sediment loading (Ghosh, 2012; Rosales-Underbrink, 2015).
Figure 5. North-south seismic profile showing the Four Brothers slump body in red. The scarp face is located just south of the slump body. Overlying water depth is 65 m. From Ghosh, 2012.

3.3.2 West Quebec Seismic Zone

Extensive landslide studies, both terrestrial and lacustrine in nature, have been conducted in the St. Lawrence Lowlands and Ottawa Valley within the Charlevoix Seismic Zone (CSZ) and the West Quebec Seismic Zone (WQSZ) of eastern Canada and northern New York. This region is within relatively close geographic proximity to Lake Champlain and has a similar late-glacial and post-glacial history, as is evidenced by the presence of fine-grained glaciomarine clay sediment deposits from the Champlain Sea in both locations (Brooks, 2013). These similarities make studies within the WQSZ valuable opportunities to observe patterns in terms of both the paleoseismic history and the behavior of the sediment in the broader region.

Terrestrial landslides within the Ottawa Valley have caused significant damage on the human timescale. Over the past 120 years, at least 23 major landslides have occurred between Ottawa and Quebec City and have caused significant infrastructure and property
damage, as well as loss of life (Duffy and Spears, 2008). In May of 1971, for example, a single landslide in Saint-Jean-Vianney, Quebec killed 31 people and destroyed a bridge and 40 homes (Duffy and Spears, 2008). In each of these landslides, failure occurred within a package of sediments known as the Leda Quick Clays, which represent the interface between Champlain Sea sediments and Lake Champlain sediments (Duffy and Spears, 2008; Brooks, 2013). As is evidenced by the large number of landslides that have occurred along this transition, these sediments are structurally sensitive and are therefore prone to mass wasting (Brooks, 2013). This weakness likely exists because the stability of the sediment is compromised when salt from the original marine environment of deposition (Champlain Sea) is flushed away by groundwater flow over time (Duffy and Spears, 2008). Important to the investigation at hand is the fact that Lake Champlain shares the same sedimentary history, and that the slumps previously studied in Lake Champlain have occurred along this same sediment boundary.

Paleolandslide studies in the Ottawa Valley, both on land and in lake sediments, suggest a seismic trigger mechanism for many of the observed events (Shilts et al., 1992; Brooks, 2013). The modern-day seismic record of eastern Canada, which includes earthquakes of at least M 7.3 in 1663 AD and M 6.0 in 1870 AD in the Charlevoix, QC region, as well as multiple earthquakes in the WQSZ as large as M 5.8 and M 6.1 on the Richter scale within the past 300 years, validates the possibility that seismic activity has also occurred on a timescale of thousands of years (Filion et al., 1991; Brooks, 2013). A study by Brooks (2013) on terrestrial landslides in the Ottawa Valley shows clusters of radiocarbon dates around ~1000 BP and ~5000 BP, where earthquakes likely occurred around each of these two dates (Brooks, 2013). Brooks proposes that both the 1 ka and 5
mass wasting events were triggered by earthquakes of M 6.1 or greater (Brooks, 2013). A separate study by Shilts et al. (1992) concludes that a series of lacustrine slumps in lakes across southern Quebec were also triggered by seismic activity based on synchronous slumps about the same age as the 1663 Charlevoix earthquake (Shilts et al., 1992). Evidence suggests that these slides also failed along the same Champlain Sea interface where failure occurred in the landslides previously discussed in both Lake Champlain and the modern-day Ottawa Valley (Shilts et al., 1992).

4. Study Location and Goals

4.1 Study Area

This investigation focuses on a study area located in Main Lake Champlain, near Essex, New York between Cannon Point to the south and the Bouquet River Delta to the north (Fig. 6). The study site has an area of 22 km² and is about 9 km south of the Four Brothers Slump and Quaker Point study areas.
Figure 6. Multibeam image showing past study areas (Four Brothers and Quaker Point study areas) as well as the present study location (Essex Study Area). See also Figure 2.
4.2 Study Objectives

This study focuses on a series of overlapping landslide deposits in an area between the Bouquet River Delta and the Essex Ferry in Lake Champlain (Fig. 6), where subaqueous surface failures have occurred along both shores of the lake. However, the western shore has the most developed landslides and debris fans. By correlating Multibeam bathymetric imagery with high resolution seismic reflection data and sediment cores (Fig. 7), the objectives of this project are to: (1) identify discrete landslides and their aerial extent and origin, (2) determine the failure dates of each of these landslides, (3) calculate the volume of the displaced material, and (4) pinpoint locations where future failure might occur and model the original geological structure of the region.

*Figure 7.* Multibeam image with all collected seismic lines overlain in white, and core locations in blue.
5. Methodology

5.1 Multibeam

Multibeam data was collected by Tom Manley using a Reson 7125 High Resolution Multibeam Echosounder aboard Middlebury College’s R/V Folger in the summer of 2015 as a part of the High Resolution Bottom Mapping of Lake Champlain project. High frequency (400 KHz) radar pulses are emitted from 512 transducers in a fan shape, with a swath width of 128°. Imagery provides centimeter resolution in the vertical direction and meter resolution in the horizontal direction, and was post-processed using CARIS HIPS and SIPS software. The resulting bathymetric map allows for the identification of regions where slope failure has occurred based on visible scarp faces and associated debris fans.

5.2 Seismic Profiles

Forty-nine seismic profiles were taken aboard Middlebury College’s research vessel, the R/V Folger, in October and November of 2015. Seismic profiles were collected using an Edgetech Model SB216 compressed high intensity radar pulse (CHIRP), with a frequency range of 2-16 kHz and a penetration range of ~100 m. Thirty-seven of the 49 lines collected were in a large grid covering the entirety of the study area, with 28 east-west oriented lines and 9 north-south oriented lines (Fig. 7). The remaining 12 lines constitute micro-surveys of two specific features of interest. Six lines constitute a north-south and east-west oriented grid over a large, unfailed sediment block in the north western corner of the study area (Fig. 7), and 6 lines make up a northwest-southeast oriented grid over an unfailed region in the southeastern section of the study area. Three
additional east-west oriented lines previously collected on August 5, 2014 were incorporated into the dataset as well. All seismic profiles were processed and key reflectors were digitized using Chesapeake Bay Technology’s SonarWizMap6. These reflectors include boundaries between bedrock, Lake Vermont sediment, Champlain Sea sediment, sediment from the transition between Champlain Sea and Lake Champlain, Lake Champlain sediment (Fig. 8). These digitized surfaces were then imported into earthVisions, a modeling and mapping software that uses digitized reflectors to best characterize the regional sedimentary layering in a fully 3-D renditioned model.
Figure 8. East-west oriented seismic line showing the seven reflectors that were digitized in each seismic profile, as well as the acoustic units that are delineated by each.
5.3 Sediment Cores and Core Processing

5.3.1 Core Collection

Seismic data was used to strategically select target locations for four piston cores. Cores were collected using a Benthos 3-inch diameter piston corer aboard the R/V Folger. Two core locations (Core 2 and Core 3) were chosen with the intention of penetrating only undisturbed Lake Champlain sediments for the purpose of determining sedimentation rates (Fig. 7). It was deemed necessary to independently derive sedimentation rates for both the east and west sides of the study area because the presence of the Bouquet River Delta to the north of the west side of the study area suggests that sediment flux is likely higher on the west side than the east, and that the two likely have different sedimentation rates. Two additional sites (Core 1 and Core 4) were selected with the intention of penetrating landslide failure surface material in order to determine the date of failure (Fig. 7). However, only Core 2 and Core 3 were successfully extracted. Retrieved cores were capped and stored at 4°C until ready for further processing.
Core 1 was attempted at the location of 44 19.6794°N and 73 20.0969°W on the west side of the study area, however the core failed, inverting the core fingers and shattering the core liner, the combination of which led to no sediment being captured (Fig. 10). This failure was likely caused by penetration into extremely dense clay, which
could have been either Champlain Sea or Lake Vermont sediment, but based on the acoustics of the seismic data, it was most likely Champlain Sea sediment (Fig. 9). Champlain Sea sediment is denser than Lake Champlain sediment because it has been compacted and water has been driven out of pore space due to overburden weight. This is increasingly true as depth increases. Core 4 was attempted at the location of 44 18.8888 N and 73 18.6639 W on the east side of the study area and also failed to capture any sediment due to the penetration of very dense clay with repaired but compromised piston core fingers.

![Figure 10](image.png)

**Figure 10.** After failed coring attempts, core fingers were inverted (left) and the core liner was shattered (right).

Cores were processed for a suite of physical properties. The cores were measured at room temperature for magnetic susceptibility (MS) at a 2 cm interval using a Bartington MS2C Core Logging Sensor with a 45 mm diameter loop. MS measures variations in the ferromagnetic mineral content of the sediment, which can be used to correlate between sediment cores (Freed et al., 1975). Cores were then split open and visual observations of grain size and sediment structures, as well as observations of color.
based on the Munsell color chart, were systematically recorded.

Samples with a volume of about 4.0 cc were taken from both cores at a 1 cm interval for the top 10 cm and every 2 cm thereafter. Samples were then weighed and dried in an oven at 105°C for 24 hours and then weighed again. Both sets of weights were used to calculate porosity and saturated bulk density. Each sample was then deflocculated using a 3% Calgon solution and analyzed for grain size using a Horiba LA-920 laser diffraction particle size analyzer.

5.3.2 Dating

Additional samples of about 4.0 cm³ were extracted at an interval of every centimeter for the top 10 cm of each core and every 2 cm from 10 – 25 cm and sent to Dr. Richard Bopp at Rensselaer Polytechnic Institute for $^{210}\text{Pb}$ and $^{137}\text{Cs}$ dating. Excess concentrations of $^{210}\text{Pb}$ are found in lake sediments as a result of the use of tetraethyl lead in leaded gasoline between the years of 1920 and 1970 (Mecray, 2000). Excess concentrations of the $^{137}\text{Cs}$ radionuclide are present due to the testing of nuclear weapons, beginning in 1954 and with a peak fall-out in 1963 (Dawson, 2008). These radionuclide signatures are both well documented in Lake Champlain and by measuring the concentration of each isotope at each of the sampled intervals within the cores, the stratigraphic depth of peak fall-outs can be determined; then, since the timing of the fall-outs is known, sedimentation rate can be calculated based on the amount sediment that has accumulated above the fall-out.

During the core splitting process, datable course organic material (e.g. plant parts)
was removed from the core and stored. A total of five organic samples – three from Core 3 and two from Core 2 – were sent to Woods Hole Oceanographic Institute for $^{14}$C dating, which determined the age of organic material based on the known half-life of the radioactive decay of organic carbon. These ages correspond to the depth within the core at which each sample was found and can be used to improve sedimentation rates and account for the possibility that sedimentation rate is not consistent throughout the entire period of Lake Champlain deposition. The samples analyzed from Core 3 were a pine needle found at a depth of 35 cm, a stem found at a depth of 235 cm, and a stem found at a depth of 313 cm. Samples analyzed from Core 2 were a stem found at a depth of 100 cm and a stem found at a depth of 158 cm.

The appearance of trace metals in lake sediments at the onset of the Industrial Revolution around 1800 is well constrained in the Champlain Valley (Mecray et al., 2001). The time at which individual elements appear varies slightly and Zn, for example, becomes elevated above background levels in 1835 (Mecray et al., 2001). Therefore, the depth in the core at which Zn concentration begins to increase represents the year 1835, and sedimentation rate can be refined based on how much sediment has accumulated since that spike. For all samples from the top 35 cm of both Core 3 and Core 2, X-ray fluorescence (XRF) analysis was performed to measure trace metal concentrations. To prepare for analysis, all samples were ignited in a LECO oven and were then combined with lithium metaborate and fluxed in a Claisse LeNeo Fluxer. Analysis was conducted for the elements Na, Mg, Al, Si, P, K, Ca, Ti, Cr, Mn, Fe, Ni, Cu, Zn, Sr, Y, Zr, Nb, Pb, and Th on a Thermo Scientific ARL Quant-X EDXRF Analyzer.
6. Results

6.1 Multibeam

From Multibeam imagery, four debris flows were identified on the west side of the study area, each with an associated scarp face and debris fan protruding out into the central region of the lake (Fig. 11). Four sections of unfailed shelf can also be identified on the west side (Fig. 11). On the east side of the study area, four debris flows and four unfailed regions can potentially be identified (Fig. 11). However, scarp faces and debris fans are much better defined on the west side of the study area, and therefore the analysis of this dataset will focus largely on the western features.

Figure 11. Multibeam image labeled with each of the features identified in the region. “W” stands for west side, while “E” stands for east side, and “S” denotes a failed area while “B” denotes an area where failure has not occurred.
Once identified, a naming scheme was developed for the purpose of referencing each feature. “W” signifies the feature is located on the west side of the study area, while “E” implies that the feature is located on the east side. “S” stands for “slump” and marks locations where failure has occurred, while “B” stands for “block” and marks regions that have not failed. WS1 is the northernmost debris flow on the west side of the study area, where failure initiated from a scarp face on the western shelf and flowed toward the center of the lake, out and around either side of WB1, which is an intact, unfailed section of shelf with dimensions of about 400 m by 300 m and a thickness of about 10 m (Fig. 11). WB2 is a small, ~50-75 m wide unfailed section of shelf that marks the southernmost extent of WS1 (Fig. 11). WS2 is the failed section of shelf with a scarp face that spans ~0.5 km from northern to southern extent and is bounded by WB2 to the north and WB3 to the south (Fig. 11), where WB3 is a narrow strip of unfailed shelf (~100 m wide) protruding from the shore of the lake toward the center (Fig. 11). Continuing to the south, WS3 is another region where a debris flow has occurred, with a scarp face that is ~1 km long from northern to southern extent. It is identifiable by its discernable scarp face and by the presence of a corresponding debris fan in the central region of the lake (Fig. 11). WB4 is an unfailed region just south of WS3 that is about 0.7 km long from northern to southern extent. It can be identified in multibeam imagery by its high relief as compared to the surrounding failed regions. The southernmost identified feature on the west side of the study area is WS4, a failed region with a scarp face about 0.7 km and a well-defined corresponding debris fan in the central region of the lake (Fig. 11).

On the east side of the study area, the northernmost feature (ES1) is a region
where failure had occurred, with a scarp face ~0.8 km long. It is identifiable based on the low relief of the area and the associated debris flow in the central region of the lake (Fig. 11). It is bounded to the south by EB1, which is a section about 0.75 km long where failure has not occurred, with the exception of a small slump about 200 m wide in the center of the unfailed region. This small slump will be referred to as ES2 (Fig. 11). ES3 is a narrow section of failed shelf, with a scarp face ~200 m long from northern to southern extent, located just south of EB1 (Fig. 11). The northern extent of ES4 is just south of ES3, and the scarp face extends ~1.5 km further to the south. The scarp face of ES4 is easily identifiable, and is located further to the west than the debris flows identified to the north. EB2 is the ~450 m wide region of unfailed shelf located to the east of ES4, EB3 is an unfailed region just west of the southern part of ES4. It is about 300 m wide in the east-west direction with a scarp face about 250 m long in the north-south direction (Fig. 11).

6.2 Seismic Profiles

Key reflectors have been digitized and color-coded to understand the sedimentary and tectonic history of the study region. Seismic profiles enable the analysis of sub-bottom sediment, which in combination with multibeam data, can be used to determine the history of failure. From the seismic profiles collected, six distinct acoustic units as well as two debris flow deposits, separated by seven key reflectors, can be identified (Fig. 8; 12-13). The uppermost reflector, which is shown in red in (Fig. 8; 12-13), indicates the sediment-water interface. The green reflector is located beneath the sediment-water interface and it is present in all locations where slope failure has occurred recently, but is
not ubiquitously present throughout the study area, including in locations where slope failure has not occurred. The acoustic unit between the red and green reflectors is acoustic unit 1, which consists of sediment from the Lake Champlain period that has accumulated since slope failure occurred (Table 1).

Table 1. Summary of acoustic units and their corresponding sediment types.

<table>
<thead>
<tr>
<th>Acoustic Unit</th>
<th>Sediment Type (Based on cores and acoustic character)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Lake Champlain sediment accumulated since slope failure (red reflector)</td>
</tr>
<tr>
<td>2</td>
<td>Lake Champlain sediment (green reflector)</td>
</tr>
<tr>
<td>3</td>
<td>Champlain Sea – Lake Champlain transition sediment (blue reflector)</td>
</tr>
<tr>
<td>4</td>
<td>Champlain Sea sediment (orange reflector)</td>
</tr>
<tr>
<td>5</td>
<td>Lake Vermont sediment (pink reflector)</td>
</tr>
<tr>
<td>6</td>
<td>Bedrock (yellow reflector)</td>
</tr>
</tbody>
</table>

Acoustic unit 2 is bounded by the green reflector and the navy blue reflector and also consists of Lake Champlain sediments. It is not ubiquitously present throughout the study area, and is missing in locations where slope failure has occurred. It is also indistinguishable from unit 1 in some locations where slope failure has not occurred, and in these locations is bounded by the red reflector and the navy blue reflector. Acoustic unit 3 is bounded by the navy blue and orange reflectors and contains sediment from the transition between the Champlain Sea and Lake Champlain sedimentation periods. It is
only present in select unfailed regions of the study area. In locations where unit 3 is missing, the green reflector and the orange reflector bound unit 2. In locations where slope failure has occurred, both acoustic units 2 and 3 are missing, and the orange reflector coincides with the green reflector (Figs. 8; 12-13).

![Image](image.png)

**Figure 12.** Section of the east-west oriented seismic profile number 6, showing a section of WS1, where the green reflector and the orange reflector overlap because acoustic units 2 and 3 were displaced during slope failure.

The orange reflector and the pink reflector bound acoustic unit 4, which contains sediment deposited during the Champlain Sea period. Acoustic unit 5 is made up of sediment deposited during the Lake Vermont Period and is bounded by the pink reflector and the yellow reflector. Acoustic unit 6 is basement rock and is located beneath the yellow reflector. Acoustic units and the sediment types they represent are summarized in
Acoustic units 2, 3, 4, 5 and 6 are visible on the slopes of the lake within the study area, but not in the center of the lake. The deep, central section of the study area is dominated by acoustically transparent debris flow deposits that are overlain by acoustic unit 1 (Figs. 12-13). The light blue reflector is not ubiquitously present throughout the study area, but when it is present in seismic lines where the navy reflector is also present, the two reflectors line up with one another. However, the light blue reflector bounds debris flow deposits and is found in central sections of the study area, whereas the navy reflector bounds undisturbed, unfailed masses of sediment in sections of the slope of the study area. In the central region of the study area, the acoustic unit bounded by the green reflector and the light blue reflector is debris flow deposit 1. The acoustic unit beneath the light blue reflector is debris flow deposit 2. The acoustic transparency of both debris flow units indicates that structural integrity was not maintained when slope failure occurred. It is important to note that while it is possible to identify the presence of these debris flow deposits, it is not possible to determine where this sediment originated from based on the seismic data. The individual debris fans associated with each region of failure that can be seen in the multibeam imagery are not distinguishable from one another in the seismic data (Fig. 13; Table 1). Stratigraphy below debris flow deposit 2 is indiscernible due to limitations of the seismic penetration energy.
In north-south lines along both the eastern and western sides of the study area, slope failures can be identified where acoustic unit 1 directly overlies acoustic unit 4. On the west side of the study area, the same four failed regions identified in multibeam...
imagery can also be identified in seismic imagery in this way (Fig. 14). The four intact blocks identified in multibeam imagery can also be identified in demonstrative north-south seismic lines as regions where a full sequence of acoustic units is present in between failed sections. On the east side of the study area, the same strategies were applied to identify failed and unfailed regions, however the seismic imagery is much less straightforward and interpretations are much less conclusive.
Figure 14. North-south oriented seismic profile 32, with failed and unfailed regions labeled. Black arrows represent the thickness of sediment that has accumulated since failure. Scale bar in bottom seismic profile applies to all seismic images.
A 3-dimensional model of stratigraphy throughout the study area helps to illustrate the spatial relationship between identified acoustic units (Fig. 15). For example, it shows how Transition Period sediments (blue unit) are not present throughout the entire study area, as well as how the reflector separating Champlain Sea sediments from Transition Period sediments in sloped regions is continuous with the reflector marking the lower boundary of the debris flow deposits in the central region of the basin (Fig. 15b). On the western side of the study area, WB3 is the only region where Transition Period sediments are present, and the blue unit in the central part of the basin represents the debris flow deposits. The model also approximates the stratigraphy of the central region of the basin, where depth of seismic penetration limits the available data, by extrapolating based on more complete data from surrounding regions. This extrapolation is not perfect, but helps to illustrate that the failure surface is continuous throughout the study region (Fig. 15c).
Figure 15. 3 Dimensional model showing the sedimentary structure of the whole study area (a), the discontinuous nature of Transition Period sediment (b), and the continuous nature of the failure surface (c). Color scheme matches that previously defined for acoustic reflectors.

6.3 Piston Cores

In this study, 4 three-inch diameter piston cores were attempted at strategically selected locations based on features visible in seismic images, but only two (Core 2 and Core 3) were successfully obtained. Target locations for Core 1 and Core 4, both of which failed, were selected with the intention of penetrating the failure surfaces of WS1 and ES4, respectively, in order to estimate approximate timing of failure (Fig. 16)
6.3.1 Core 3

Core 3 was 361 cm long and was extracted from the west side of the study area, at the location 44 18.4750°N and 73 20.2649°W and with an overlying water depth of 75 m (Fig. 17). This core was located on WB3 and based on seismic stratigraphy, only penetrated undisturbed laminated Lake Champlain sediments (Fig. 17). This location was selected in order to determine a sedimentation rate for the west side of the study area.
Visual analysis of the split core reveals little variation in the appearance of the sediment throughout the length of the core. According to the Munsell color chart, the sediment is olive gray (5Y 4/1) throughout, with clusters of faint black laminations visible throughout the length of the core. These laminations are most likely hydrotolites, a form of hydrated iron sulfide (FeS•nH₂O) that forms early in the process of diagenesis (Reddy and DeLaune, 2008; Ghosh, 2012). Laminations are approximately 1 to 4 mm wide and occur in clusters of about 2-3 laminations every centimeter.
Magnetic susceptibility (MS) data for Core 3 shows a large drop off in magnetic material at a depth of about 25 cm, and from there increases gradually with increasing depth (Fig. 18). From 0-25 cm, MS fluctuates between 25x10^{-5} and 30x10^{-5} SI, and then drops down to 17.5x10^{-5} SI by the depth of 44 cm. This depth is representative of about the year 1760, when colonists first began to clear-cut the area (P. Manley, per. comm.). From this point on, as depth within the core increases, MS also increases gradually, reaching values in the range of 20x10^{-5} to 27.5x10^{-5} SI by a depth of about 350 cm.

Porosity is fairly consistent and fluctuates between 85.0 and 89.5% throughout the entire core, and the small decrease in percent porosity that is observed downcore is consistent with an increase in overburden (Fig. 18). This is also observed in an increase in density, which is due to increased pressure and compaction with increasing depth rather than the penetration of a distinct acoustic unit. Saturated bulk density is the inverse of porosity and is also consistent throughout the length of the core, with values that fall in the tight range of 1.2 to 1.3 g/cm³. Grain size is relatively constant throughout the core, with the exception an increase in diameter at a depth of 145-150 cm (Fig. 18). The mean grain size is 10.48 ± 9.42 µm, with a median grain size of 8.57 µm.
Figure 18. Physical properties, magnetic susceptibility, and grain size frequency distribution for Core 3.
6.3.2 Core 2

Core 2 is 232 cm long and was extracted from the east side of the study area, at the location 44 19.8818°N and 73 18.6560°W with an overlying water depth of 80 m (Fig. 17). This core is located on EB1 and based on seismic stratigraphy only penetrated undisturbed Lake Champlain sediments (Fig. 17). This location was selected in order to determine a sedimentation rate for the east side of the study area.

Visual analysis upon splitting Core 2 also reveals little variation in the sediment throughout the length of the core. According to the Munsell color chart, like Core 3, the sediment in Core 2 is olive gray (5Y 4/1) throughout, with very tightly spaced hydrotolites in the top 10 cm and then about 1 cm spacing throughout the rest of the core.

MS is about 16x10⁻⁵ SI for the top 4 cm of Core 2, and then drops to 3.1x10⁻⁵ SI by the depth of 14 cm (Fig. 19). Between the depths of 14 cm and the bottom of the core at 232 cm, MS fluctuates within the range of 11.7x10⁻⁵ to 16x10⁻⁵ SI, with values falling slightly closer to the upper end of the range with increasing depth in the core. Porosity in Core 2 is also relatively consistent throughout the length of the core, fluctuating within the range of 85.0 to 89.0%. Saturated bulk density is also consistent throughout the length of the core, with values that fall in the range of 1.2 to 1.3 g/cm³. Grain size is relatively constant throughout the core, except for a slight shift in frequency around a depth of 20 cm (Fig. 19). The mean grain size is 10.11 ± 8.6 µm and the median is 8.52 µm.
Figure 19. Physical properties, magnetic susceptibility, and grain size frequency distribution for Core 2.
6.4 Dating

6.4.1 $^{137}$Cs and $^{210}$Pb Dating

In Core 3, which was extracted from WB3 on the west side of the study area, the 1963 peak associated with fallout from nuclear weapon testing is located at a depth of 6 cm, roughly double the depth at which this peak was found in core 2 (Fig. 20). This is likely because the west side receives a greater sediment flux than the east side due to its location just south of the Bouqet River delta (Fig. 6). Samples taken at depths of 3, 4, 6 and 7 cm have $^{137}$Cs values of 178, 104, 124 and 15 pCi/Kg respectively. The 15 pCi/Kg value indicates that sediment at this depth and below was deposited before the peak fallout of 1963. The full dataset can be found in Appendix 1. Assuming that the depth of 6 cm represents the 1963 peak, a sedimentation rate for the west side of the study area was calculated as follows:

$$6 \text{ cm} \div (2015-1963) \text{ years} = 0.12 \text{ cm/year}$$
Core 2, extracted from EB1 on the east side of the study area, had a $^{137}$Cs signature detectable to a depth of 2-3 cm (Fig. 20). The full dataset can be found in Appendix 1. Assuming that a depth of 3 cm represents the 1963 peak in Core 2, a sedimentation rate for the east side of the study area was calculated as follows:

$$3 \text{ cm} \div (2015-1963) \text{ years} = 0.06 \text{ cm/year}$$

This is within the range of other sedimentation rates that have previously been calculated in different regions of Lake Champlain, including a sedimentation rate of 0.07 cm/yr as calculated by Ghosh (2012) near Four Brothers Island, and 0.083 cm/yr as calculated by Rosales-Underbrink (2015) near Quaker Point. Other documented sedimentation rates for the Lake Champlain period range from 0.05 cm/yr near Barber Point in Folger’s Trough in the southern section of the lake (Cronin et al., 2008) to 0.15 cm/yr in the main lake (Hunt, 1976).
6.4.2: XRF Analysis of Heavy Metal Concentrations

To augment the $^{210}$Pb and $^{137}$Cs dating, XRF was conducted for a suite of elements including heavy metals in order to identify the first evidence of the Industrial Revolution in lake sediments. Of the heavy metals for which XRF analysis was conducted, zinc best displays the characteristic increase above background levels at the beginning of the Industrial Revolution. According to Mecray et al. (2001), heavy metal concentrations (Cu, Zn, Cr, Pb, Cd, Ag) in Lake Champlain sediment show little variation in sediments older than the onset of the Industrial Revolution, but concentrations begin to increase above background in the early 1800s (Mecray et al., 2001). Specifically, Zn first begins to gradually increase above background levels around the year 1835, which can be attributed to fossil fuel and coal combustion as well as the fabric treatment process used in the textile industry (Mecray et al., 2000). In Core 3 (west side), this increase above background level occurs between the depths of 15-17 cm (Fig. 21). This means that a depth of 15 or 17 cm represents the year 1835, and a range of sedimentation rates for the west side of the study area can be calculated as follows:

\[
\frac{15 \text{ cm}}{(2015-1835) \text{ years}} = 0.083 \text{ cm/year}
\]

\[
\frac{17 \text{ cm}}{(2015-1835) \text{ years}} = 0.094 \text{ cm/year}
\]

In Core 2 (east side), the concentration of Zn increases above background level at a depth of about 13 cm, and a sedimentation rate for the east side of the study area can be calculated as follows:

\[
\frac{13 \text{ cm}}{(2015-1835) \text{ years}} = 0.072 \text{ cm/year}
\]
Complete XRF data can be found in Appendix 2.

Figure 21. XRF data showing the concentration of Zn in sediment at various core depths for Core 3 (black) and Core 2 (green).

6.4.3 \(^{14}\)C Dating

Radiocarbon dates were calibrated to calendar years before present (BP) using the CalPal software (Table 2). After applying this correction, for Core 3 samples from the west side of the study area, the samples at 35, 235 and 313 cm had ages of 134±100, 1683±46 and 1457±43 years respectively (Table 2). The slightly older age at a depth of 235 cm as compared to the sample from 313 cm may be explained by a variety of possible factors. One possibility is that sediment reworking had occurred at 235 cm, and
older organic material that was preserved elsewhere was released back into the water column and re-deposited, meaning that the carbon date and the date of sedimentation are not synchronous. Another possibility is that slightly older organic matter that was preserved in a different environment was carried into the lake by the Bouquet River during a storm or flood event. For these reasons, the 1683±46 age at 235 cm was not used in the calculation of overall sedimentation rate ranges.

For Core 2 (east side), the sample at a depth of 100 cm had an age of 7008±202 while the sample at a depth of 158 cm has an age of 1854±20 (Table 2). This suggests that the former sample is about 5000 years older than the latter, even though it was found 58 cm shallower in the core and should therefore be expected to be younger. While the cause of the anomalous age cannot be determined, it is reasonable to assume that the age of 7008±202 at 100 cm should not be included in the age model. $^{137}\text{Cs}$ and $^{210}\text{Pb}$ data shows that the sedimentation rate on the east side of the study area is about half that of the west side of the study area, and this $^{14}\text{C}$ age yields a sedimentation rate of $\sim0.01$ cm/yr, which is significantly less than half of the sedimentation rate calculated for the west side of the study area. The age of 1854±20 at a depth of 158 cm, on the other hand, yields a sedimentation rate of 0.085 cm/yr, which is close to the sedimentation rate of 0.07 cm/yr that was calculated for the Four Brothers Island region to the north of the current study area (Ghosh, 2012; Rosales-Underbrink, 2015). Full $^{14}\text{C}$ data can be found in Appendix 3.
Table 2. Summary of $^{14}$C data. Ages were calibrated to calendar years before present using the online CalPal service.

<table>
<thead>
<tr>
<th>Core</th>
<th>Sample Depth (cm)</th>
<th>Calibrated $^{14}$C age (calendric year BP)</th>
<th>Sedimentation Rate (cm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Core 3</td>
<td>35 cm</td>
<td>134 ± 100</td>
<td>0.26 cm/yr</td>
</tr>
<tr>
<td></td>
<td>235 cm</td>
<td>1683 ± 46</td>
<td>0.12 cm/yr</td>
</tr>
<tr>
<td></td>
<td>313 cm</td>
<td>1457 ± 43</td>
<td>0.17 cm/yr</td>
</tr>
<tr>
<td>Core 2</td>
<td>100 cm</td>
<td>7008 ± 202</td>
<td>0.01 cm/yr</td>
</tr>
<tr>
<td></td>
<td>158 cm</td>
<td>1854 ± 20</td>
<td>0.09 cm/yr</td>
</tr>
</tbody>
</table>

6.4.4 Overall Sedimentation Rates

Sedimentation rates were independently calculated based on $^{137}$Cs and $^{210}$Pb dating, $^{14}$C dating, and XRF data for Zn concentration. The overall sedimentation rates used to calculate failure ages on both the west and east side of the study area assume that sedimentation rate has remained constant throughout the Lake Champlain period and are based on linear regression models that include data collected through all three dating techniques (Fig. 22). The resulting range in sedimentation rates comes from the range of possible depths at which Zn spikes above background levels, as well as the inclusion and exclusion of individual $^{14}$C data points due to their discordant chronology. However, the minimum sedimentation rate for the east side of the study area was calculated by taking the arithmetic mean of all values because the $R^2$ value for the linear regression was extremely low due to the discordance in included $C^{14}$ ages (Fig. 22). For the west side of the study area, this range in sedimentation rate is 0.13-0.17 cm/yr, and for the east side of the study area it is 0.06-0.085 cm/yr (Table 3). While these ranges include all possible
variability in sedimentation rate based on collected data, the sedimentation rate for the west side of the study area is likely closer to the 0.13 cm/yr end of the range since 0.13 cm/yr best fits the expected ages of sediment accumulated above the Champlain Sea-Lake Champlain interface in places where failure has not occurred, based on previous work which suggests that this transition occurred ~9-10 ka (Cronin et al., 2008).
Figure 22. Plot showing age versus core depth for all age data, as well as linear regression plots showing the maximum and minimum sedimentation rates for both Core 3 (west side) and core 2 (east side). Minimum and maximum rates were calculated based on various interpretations of all collected age data.
Table 3. Table summarizing sedimentation rates calculated based on various dating methods, used to determine an average sedimentation rate for both the west and east sides of the study area.

<table>
<thead>
<tr>
<th>Core</th>
<th>Dating Method</th>
<th>Calculated Sedimentation Rate</th>
<th>Overall Sedimentation Rate</th>
</tr>
</thead>
<tbody>
<tr>
<td>Core 3</td>
<td>$^{137}$Cs and $^{210}$Pb Dating</td>
<td>0.12 cm/yr</td>
<td>0.13-0.17 cm/yr</td>
</tr>
<tr>
<td></td>
<td>XRF</td>
<td>0.083-0.094 cm/yr</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$^{14}$C Dating</td>
<td>0.26 cm/yr</td>
<td></td>
</tr>
<tr>
<td>Core 2</td>
<td>$^{137}$Cs and $^{210}$Pb Dating</td>
<td>0.06 cm/yr</td>
<td>0.06-0.085 cm/yr</td>
</tr>
<tr>
<td>(west side)</td>
<td>XRF</td>
<td>0.072 cm/yr</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$^{14}$C Dating</td>
<td>0.085 cm/yr</td>
<td></td>
</tr>
<tr>
<td>Core 2</td>
<td>$^{137}$Cs and $^{210}$Pb Dating</td>
<td>0.12 cm/yr</td>
<td>0.13-0.17 cm/yr</td>
</tr>
<tr>
<td>(east side)</td>
<td>XRF</td>
<td>0.083-0.094 cm/yr</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$^{14}$C Dating</td>
<td>0.26 cm/yr</td>
<td></td>
</tr>
</tbody>
</table>

7. Observations and Discussion

7.1 Core-Seismic Correlation

Seismic imagery indicates the presence of four major acoustic units which, based on previous work in the region by Hunt (1972), Cronin et al. (2008) and Dawson (2008), are indicative of three depositional periods with distinct environments since the last glacial retreat in the late Wisconsin (Fig. 4). The seismic profiles collected in the Essex study region penetrate down to bedrock at a maximum depth of 150 m below lake level. Bedrock is overlaid by Lake Vermont sediments (acoustic unit 4 in Fig. 8), which are in turn overlaid by Champlain Sea sediments (Acoustic unit 3 in Fig. 8). In some locations, sediments from the transition period between Champlain Sea and Lake Champlain are present above acoustic unit 3 (acoustic unit 2 in Fig. 8). Lake Champlain sediments make up the upper acoustic unit (acoustic unit 1 in Fig. 8). Although seismic data indicates that cores only penetrated Lake Champlain sediments, the nature of the other
units can be inferred based on the known stratigraphy of the Champlain Basin and the known acoustic signature of each unit. Bedrock constrains the overall character of the basin, and the slopes, or rises, from which debris flows originated are a product of the bedrock geometry.

The locations for both Core 3 and Core 2 were selected with the intention of only penetrating Lake Champlain sediment, and visual and physical properties of both cores confirm that only Lake Champlain sediment was penetrated. The lack of change in the visual appearance of sediment throughout the entire length of the core, including the consistent olive gray color, clay-like texture and sporadic presence of hydrotolites, also suggests that only one acoustic unit was penetrated.

The physical properties of both cores also confirms that only one acoustic unit was penetrated. The lack of variation in both porosity and saturated bulk density throughout the length of both cores indicates that the nature of the sediment is similar throughout, and that only one acoustic unit was in fact penetrated. In Core 3, porosity remains within the tight range of 85.0 to 89.5% and saturated bulk density remains within the range of 1.2 to 1.3 g/cm³. In Core 2, porosity stays within the range of 85.0 to 89.5% and saturated bulk density fluctuates only between 1.2 and 1.3 g/cm³. Furthermore, the similarity in the values for grain size, porosity and saturated bulk density between the two cores also confirms that both cores contain sediment from the same unit.

In a study using similar methodology in the Willsboro Bay region of Lake Champlain, Dawson (2008) was able to core through three units, and from those cores was able to identify Lake Champlain (unit 1), Champlain Sea (unit 2) and Lake Vermont (unit 3) sediments. The physical properties used to identify each unit are listed in Table
4. The saturated bulk density and porosity values for unit 1 in Dawson’s cores WBAY 6-3, WBAY 6-2 and WBAY 6-1B are very similar to the respective ranges for both Core 3 and Core 2 in the present study, which further confirms that the unit that was penetrated is Lake Champlain sediment.

Table 4. Table modified from Dawson (2008) presents the different physical characteristics of each of the acoustic units penetrated by the cores in her study. Red columns correspond to Lake Champlain (Unit I) sediments, blue corresponds to Champlain Sea (Unit II) and green corresponds to Lake Vermont (Unit III).

<table>
<thead>
<tr>
<th></th>
<th>WBAY 6-3</th>
<th>WBAY 6-2</th>
<th>WBAY 6-1B</th>
<th>WBAY 11-4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean Grain Size</td>
<td>Unit 1</td>
<td>Unit 2</td>
<td>Unit 1</td>
<td>Unit 2</td>
</tr>
<tr>
<td></td>
<td>4-5</td>
<td>-4</td>
<td>3-17</td>
<td>3-7</td>
</tr>
<tr>
<td>Magnetic Susceptibility</td>
<td>5-35</td>
<td>76-225</td>
<td>20-65</td>
<td>185-260</td>
</tr>
<tr>
<td>Electric Resistivity</td>
<td>6.5-7.5</td>
<td>4-8</td>
<td>5.5-7.5</td>
<td>6-7</td>
</tr>
<tr>
<td>Saturated Bulk Density</td>
<td>1.2-1.4</td>
<td>1.5-1.8</td>
<td>1.2-1.5</td>
<td>1.7-1.8</td>
</tr>
<tr>
<td>Porosity</td>
<td>77-89</td>
<td>60-72</td>
<td>75-95</td>
<td>60-65</td>
</tr>
</tbody>
</table>

7.2 Failure Surface

Seismic data shows that in locations where the slope had failed, the failure initiated along the orange reflector (Fig. 12). The only penetrable sediment in the central region of the lake is displaced and disturbed debris flow deposits, but on the topographic rises on both the east and west sides of the study area, stratigraphy can be deciphered. In the regions where stratigraphy can be deciphered, the orange reflector is continuous throughout the study area, whereas overlying reflectors are present in some locations but are missing in others (Figs. 8; 12-13). This suggests that failure occurs along the orange reflector, at the interface between acoustic units 2 and 3. Based on knowledge of the stratigraphy of the Lake Champlain basin and the acoustic signature of each unit, the
surface along which failure occurred is likely the interface between Champlain Sea sediment and the overlying transition period or Lake Champlain sediments. The similarities in physical properties between Lake Champlain sediments (unit 1) in Dawson’s (2008) cores and Cores 2 and 3 from the present study indicate that the sediment above the orange reflector is Lake Champlain sediment, and based on known stratigraphy it can then be assumed that the underlying acoustic unit is Champlain Sea sediment. Ghosh’s (2012) investigation of the Four Brothers slump also confirms that failure occurs along the interface between Champlain Sea sediments and overlying sediments. In her FB Core 1, she penetrated a slump deposit along with the underlying failure surface, and based on both seismic stratigraphy and physical properties of sediment samples, this underlying unit matches the previously documented characteristics of Champlain Sea sediment (Dawson, 2008; Ghosh, 2012).

Landslides studied in other regions of Lake Champlain have also failed along this Champlain Sea-Lake Champlain interface (Ghosh, 2012; Rosales-Underbrink, 2015). This interface is likely more prone to failure because of differences in the character of the Lake Champlain sediments above and the Champlain Sea sediments below. According to Dawson’s study, where both Lake Champlain and Champlain Sea sediments were penetrated, the Lake Champlain sediment has a density of 1.2-1.5 g/cm³, whereas the Champlain Sea sediment has a higher density of 1.5-1.8 g/cm³, making it less permeable than the overlying sediment (Table 4) (Dawson, 2008; Ghosh, 2012). As density increases, porosity decreases and the ability of water to permeate through the sediment also decreases, lubricating this interface and making it more likely to fail. This difference in density is caused by the composition of the sediment, where Lake Champlain
sediments are coarse and high in organic matter compared to predominately terrigenous nature of the Champlain Sea sediments (Ghosh, 2012).

7.3 Age of Failure

Measurement of stratigraphic thickness in seismic profiles via the measurement tool in SonarWizMap was used to estimate the depth of the failure surface beneath the sediment-water interface. Given that the orange reflector, or the Champlain Sea-Lake Champlain interface, has been identified as the failure surface, the thickness of sediment present above this reflector was systematically measured throughout all seismic lines within the failed blocks. Seismic data was then corroborated with multibeam imagery to identify the thickness measurements that correspond with each failed region. For these failed regions, the sediment present above the orange reflector has accumulated since failure occurred, and an approximate date of failure can be calculated by dividing this sediment thickness by the sedimentation rate to get the number of years it would have taken for that amount of sediment to accumulate.

On the west side of the study area, WS1 had an average accumulation of 1.2 m of sediment above the failure surface, and given a sedimentation rate of 0.13-0.17 cm/yr, an approximate failure age of about 950-1200 ya. WS2 had an average accumulation of 4.9 m of sediment above the failure surface and an approximate failure age of about 3800-5000 ya. WS3 had an average accumulation of 5.7 m of sediment with an approximate failure age of about 4400-5800 ya. Finally, WS4 had an accumulation of about 5.9 m and an approximate failure age of about 4500-5900 ya. This data is summarized in Table 5.
Table 5. Summary of the identified regions where failure has occurred on the west side of the study area and the approximate age of failure and thickness of sediment accumulation above the failure surface for each. Failure ages for previously studied failures in the Four Brothers and Quaker Point study areas are also included.

<table>
<thead>
<tr>
<th>Failure</th>
<th>Average Sediment Accumulation Since Failure</th>
<th>Age of Failure</th>
</tr>
</thead>
<tbody>
<tr>
<td>WS1</td>
<td>1.2 m n = 84</td>
<td>950-1200 ya</td>
</tr>
<tr>
<td>WS2</td>
<td>4.9 m n = 30</td>
<td>3800-5000 ya</td>
</tr>
<tr>
<td>WS3</td>
<td>5.7 m n = 38</td>
<td>4400-5800 ya</td>
</tr>
<tr>
<td>WS4</td>
<td>5.9 m n = 21</td>
<td>4500-5900 ya</td>
</tr>
<tr>
<td>Four Brothers</td>
<td>-</td>
<td>4550 ya</td>
</tr>
<tr>
<td>Quaker Point North</td>
<td>-</td>
<td>5130-5520 ya</td>
</tr>
<tr>
<td>Quaker Point Central</td>
<td>-</td>
<td>4630-4985 ya</td>
</tr>
</tbody>
</table>

The calculated failure age ranges for WS2, WS3 and WS4 overlap with each other from 4500-5200 ya. Previously calculated failure ages of the Four Brothers Slump (4550 ya) and the Quaker Point Central (4630-4985 ya) and Northern (5130-5520 ya) debris flows also fall within this age range (Fig. 23). This means that a total of six coeval slope failures between 4500-5200 ya have now been identified in Lake Champlain. The synchronicity of these failures also suggests that they were likely triggered by the same event.
Figure 23. Plot showing failure ages of coeval failure events in Lake Champlain and the volume of sediment displaced for each. Essex region failures are shown in blue, Quaker Point failures are purple, and Four Brothers Slump is green. The gray box highlights the 4500-5200 ya time interval.

On sections of the shelf that have not failed, the sediment that has accumulated has a thickness that corresponds to the amount of time that has passed since the onset of post-Champlain Sea sedimentation. Work done by Cronin et al. (2008) suggests that Champlain Sea sedimentation occurred from ~13 ka – 9 ka, and therefore the sediment accumulated above the upper Champlain Sea reflector in unfailed regions should be representative of about 9000 years of sedimentation. This supports the interpretation that these regions have not failed and also supports the validity of the calculated sedimentation rates in predicting the amount of time sediment has been accumulating. For all unfailed areas, the sedimentation rate of 0.13 cm/yr gives a much more reasonable age than the sedimentation rate of 0.17 cm/yr, which yields ages of sediment accumulation that represents less time than an unfailed region should (Table 6).
Table 6. Summary of west side unfailed blocks and the amount of time represented by the sediment accumulations above the interface between Champlain Sea sediments and overlying sediment. Ages were calculated by dividing the stratigraphic thickness above the Champlain Sea interface by both end members of the calculated sedimentation rate range of 0.13-0.17 cm/yr.

<table>
<thead>
<tr>
<th>Block</th>
<th>Age of Sediment Accumulation</th>
</tr>
</thead>
<tbody>
<tr>
<td>WB1</td>
<td>5600-8200 ya</td>
</tr>
<tr>
<td>WB2</td>
<td>5450-7100 ya</td>
</tr>
<tr>
<td>WB3</td>
<td>8250-9150 ya</td>
</tr>
<tr>
<td>WB4</td>
<td>7000-10150 ya</td>
</tr>
</tbody>
</table>

7.4 Volumes of Failed Material

Three-dimensional modeling was used to calculate the volume of sediment displaced by each of the identified failures on the west side of the study area. By identifying scarp faces and estimating the extent of the original shelf, polygons were generated to mimic the pre-failure structure, and the volume of each of these polygons was calculated (Fig. 24) (T.O. Manley, per. Comm.).
According to this analysis, a total of $4.62 \times 10^7$ m$^3$ of sediment was displaced from the west side of the study area due to mass wasting events. During WS1 alone, a volume of $3.3 \times 10^7$ m$^3$ of sediment was displaced, which is over an order of magnitude greater than the volume of sediment displaced by any of the previously studied mass wasting events on Lake Champlain (Table 7; Fig. 25). WS2, WS3 and WS4 were all much smaller and much more similar in size to one another, displacing $3.6 \times 10^6$, $5.6 \times 10^6$, and $4.0 \times 10^6$ m$^3$ of sediment, respectively (Fig. 25). These volumes are also much more similar to the failure events previously studied on Lake Champlain. Considering that WS2, WS3 and WS4 all failed 4500-5000 ya, a volume of $1.32 \times 10^7$ m$^3$ of sediment would have been displaced at once in the Essex region of the lake if all of these failures
occurred simultaneously. Looking at Lake Champlain as a whole, if all of the coeval failures during this time period (Four Brothers Slump, Quaker Point Northern Debris Flow, Quaker Point Central Debris Flow, WS2, WS3, and WS4) occurred at the same time, a total of $2.65 \times 10^7$ m³ of sediment would have been displaced at once.

**Figure 25.** Plot showing failure ages on the x-axis and volume of sediment displaced on the y-axis for all identified failures on the west side of the Essex study region.
Table 7. Volumes of sediment displaced by each mass wasting event that has been studied in Lake Champlain.

<table>
<thead>
<tr>
<th>Failure</th>
<th>Volume of Sediment Displaced (10^6 m^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Diamond Island Slump</td>
<td>0.1</td>
</tr>
<tr>
<td>Whallon Bay Slump</td>
<td>4.0</td>
</tr>
<tr>
<td>Folger Trough Slump</td>
<td>0.23</td>
</tr>
<tr>
<td>Four Brothers Slump</td>
<td>0.055</td>
</tr>
<tr>
<td>Quaker Point Northern Debris Flow</td>
<td>9.5</td>
</tr>
<tr>
<td>Quaker Point Central Debris Flow</td>
<td>3.2</td>
</tr>
<tr>
<td>Quaker Point Southern Debris Flows</td>
<td>3.7</td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>WS1</td>
<td>33.0</td>
</tr>
<tr>
<td>WS2</td>
<td>3.6</td>
</tr>
<tr>
<td>WS3</td>
<td>5.6</td>
</tr>
<tr>
<td>WS4</td>
<td>4.0</td>
</tr>
</tbody>
</table>

7.5 Trigger Events

The possible trigger mechanisms outlined in the introduction for lacustrine landslides include wave action, storm surge, rapid sediment overloading, and seismic activity. However, with an overlying water depth of 25 m or greater, the study area is located below wave base, and Lake Champlain does not experience large storm surges (T.O. Manley, per. comm.), meaning that these trigger mechanisms were not likely the cause of the observed failures. Sediment overloading could be a possible trigger
mechanism for the failures on the west side of the study area because of the increased sediment flux due to the Bouquet River, especially for the northernmost failure (WS1), which is located in the closest proximity to the delta and did not fail at the same time as any previously studied mass wasting events on Lake Champlain. However, even though other failures ~1000 ya have not been identified on Lake Champlain, Brooks et al. (2013a) has recorded evidence of coeval terrestrial landslides in Ottawa, Canada. Coeval failure is a key indicator of seismically triggered failure, and it is therefore also possible that a seismic trigger mechanism may have been responsible for WS1. There is strong evidence for the seismically induced failure of WS2, WS3, and WS4 because not only are they roughly coeval with one another, but they are also coeval with the Four Brothers Slump and Quaker Point Northern and Central Debris Flows on Lake Champlain, as well as another set of terrestrial landslides in Ottawa studied by Brooks et al. (20113a), for which a seismic trigger mechanism has been previously established (Fig. 26) (Ghosh, 2012; Brooks et al., 2013a; Brooks et al., 2013b; Rosales-Underbrink, 2015).

Seismically induced slope failure is thought to require an earthquake of M 6.1 or higher (Brooks, 2013a). While there is generally not a lot of seismic activity on the eastern margin on North America, the Charlevoix Seismic Zone (CSZ) of Quebec has been identified as the most seismically active region of eastern North America and occasionally experiences large earthquakes, such as a M 7.4 earthquake in the year 1663 AD (Adams and Basham, 1989; Shilts et al., 1992). The nearby WQSZ has also had a notable seismic history in modern times, including several M 5.0 or greater earthquakes (Brooks, 2013a).
Figure 26. Regional summary of the seismic events recorded by slope failures. Lake Champlain data is overlaid on data from Brooks (2013). Blue boxes mark failures recorded on the west side of the current study area, and the yellow box shows the age range of previously dated failures on Lake Champlain (Four Brothers Slump and Quaker Point Debris Flows).

These historic earthquakes are further support for the likelihood that seismic events of the magnitude necessary to induce slope failure occurred in this region on a longer timespan as well (Lamontagne, 2003). Work done by Brooks (2013a) in the WQSZ shows evidence for seismic events of at least M 6.1 around both 1000 ya and 5000 ya based on C^{14} dates from terrestrial landslides that occurred along the same Champlain Sea interface that has been identified as the failure surface in Lake Champlain.
The WS1 failure in the present study occurred 950-1200 ya. This is the first time a mass wasting event of this age has been identified on Lake Champlain, but the robust cluster of failures around 1000 ya in Brooks’ dataset (Fig. 26) shows evidence of coeval failure on a regional scale (Brooks, 2013a). Therefore, it is likely that a seismic event in the region about 1000 ya could have triggered all of these ~1 ka failures.

The failure age ranges for WS2, WS3 and WS4 overlap from 4500-5200 ya (Fig. 23) and are coeval with the Four Brothers Slump and Quaker Point Central Debris Flow on Lake Champlain (Ghosh, 2012; Rosales-Underbrink, 2015). This time interval is also coeval with the radiocarbon date ranges for a significant subset of terrestrial landslides that Brooks et al. (2015) has studied in Ottawa. Brooks et al. (2015) has attributed these failures with an earthquake of about M 6.4 that occurred in the WQSZ, approximately 5000 ya, and it is likely that this same earthquake triggered the six 4500-5200 ya failures that have been identified on Lake Champlain as well (Brooks et al., 2015).
8. Conclusions and Future Work

Through correlations between multibeam and seismic data, this study has identified a total of 8 regions where slope failure has occurred, four of which are on the west side of the study area and four of which are on the east. Sediment accumulates at a rate of about 0.13-0.17 cm/yr on the west side of the study area and 0.06-0.09 cm/yr on the east side. The failures on the west side of the study area are much more clearly defined by their pronounced scarp faces and debris fans. Measurement of seismic thickness accompanied by a simple calculation using the appropriate sedimentation rate yielded approximate dates of failure for each of these western failures.

A key result of this study is the furthered evidence of coeval failure on Lake Champlain ~4500-5200 ya, which supports the conclusion that a large seismic event occurred within the region about 5000 ya. The identification of a failure event ~950-1200 ya is the first of this age to be studied on Lake Champlain, however regional evidence of terrestrial landslides with synchronous ages is indicative of a seismic trigger mechanism (Brooks et al., 2015).

In the future, a whole lake seismic study should be conducted in order to identify all failed regions and determine whether other mass wasting events occurred within the lake ~1000 ya, which would help to confirm the hypothesized seismic trigger mechanism for WS1, as well as to identify additional failures that occurred ~4500-5200 ya and to explore possible evidence for additional seismic events. Such a study would provide the dataset necessary to investigate a potential recurrence interval on seismic activity in the region.
Future work should also include numerical modeling of potential lake tsunamis generated by the identified slope failures using Cornell University’s COMCOT modeling software package. It would be particularly interesting to generate a model that simulates the simultaneous failure of all of the ~5000 ya failure events that have been identified in Lake Champlain. A model generated by Manley et al. (2015) shows that the Four Brothers Slump and the Quaker Point central and northern debris flows caused an 8 m wave height that impacted Colchester Point three minutes after the initiation of failure (Manley et al., 2015); this model should be expanded to include the present dataset.

It would also be advisable to generate a model to assess the lake tsunamis that might be generated were one of the intact regions to fail in the future. Were another earthquake to occur in the region in the future, this would have significant social significance because settlement along the shores of Lake Champlain would be affected.
9. References


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