POSTGLACIAL SEISMICITY IN ONTARIO-QUEBEC

POSTGLACIAL SEISMICITY IN ONTARIO-QUEBEC DETERMINED THROUGH ANALYSIS OF DEFORMATION STRUCTURES IN LAKE SEDIMENTS

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TITLE: Postglacial seismicity in Ontario-Quebec determined through analysis of deformation structures in lake sediments

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ABSTRACT

Eastern North America experiences large intracratonic earthquakes that are not well understood but pose a risk to urban centers and other infrastructure. Compilation of regional earthquake epicentres for south-central Ontario and western Quebec demonstrate a close association with sutures and failed rifts (the St. Lawrence Rift) recording the formation and breakup respectively of successive supercontinents Rodinia and Pangea. Where seismic potential could be underestimated through lack of historical seismicity or where little is known about active faults, lake deposits can provide a valuable record of past seismic shaking events in the form of sediment deformation structures (i.e. 'seismites'). In central Canada, the lacustrine seismographic record began approximately 10,000 years before present with the retreat of the Laurentide Ice Sheet, older records having been removed by glacial erosion. Most bedrock lake basins are structurally-controlled and underlain by the same Precambrian basement structures (shear zones, terrane boundaries and other lineaments) implicated as the source of ongoing mid-plate earthquake activity.

High resolution seismo-stratigraphic data presented here supports the model that ongoing mid-plate earthquake activity is a consequence of brittle deformation of the upper crust of the North American plate. Such activity appears to have been greatest during deglaciation but continues today. The detailed geophysical and sedimentary studies, as shown here, have major societal relevance in areas of eastern North America affected by intraplate earthquakes. The recognition and mapping of earthquake related features in lakes for seismic risk analysis is a means of constraining seismic recurrence intervals and more realistically assess seismic risk across the populated area of Ontario and Quebec where events occur on time scales much longer than recorded history.

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For this paper, all data collection in the field (seismic subbottom, bathymetry, sidescan and magnetometer data) was accomplished by M. Doughty. Dr. N. Eyles was in charge of logistics and navigation. Digitization of additional bathymetric information (not collected in the field) and draft interpolation was performed by L. Daurio (under direction); final interpolation and all GIS work were completed by M. Doughty. Processing of seismic subbottom data was conducted by M. Doughty; final interpretation of this imagery (and completion of these figures) by all authors. The remaining figures were prepared by M. Doughty. Manuscript preparation was by M. Doughty and Dr. N. Eyles.

Chapter 3 - Ongoing neotectonic activity in the Timiskaming-Kipaawa Area of Ontario and Quebec

For this paper, all data collection in the field (seismic subbottom, sidescan and magnetometer data) was accomplished by M. Doughty. Dr. N. Eyles was in charge of logistics and navigation. Multi-beam data was provided by the Canadian Hydrographic Survey (2005). Extraneous datasets (e.g. remotely sensed imagery and digital elevation models) were assembled and manipulated by M. Doughty. All GIS work, seismic subbottom processing, sidescan imagery compilation and figure preparation was by M. Doughty. Manuscript preparation was by M. Doughty and Dr. N. Eyles.

Chapter 4 - Recent neotectonic faulting of lake floor sediments in an intracratonic rift and the implications for regional seismicity: Timiskaming Graben, Canada

For this paper, all data collection in the field (seismic subbottom, sidescan and magnetometer data) was accomplished by M. Doughty. Dr. N. Eyles was in charge of logistics and navigation; additional field assistance was available during one field season. Multi-beam data was provided by the Canadian Hydrographic Survey (2005). Extraneous datasets (e.g. remotely sensed imagery and digital elevation models) were assembled and manipulated by M. Doughty. All GIS work, seismic subbottom processing, sidescan imagery compilation and figure preparation was by M. Doughty. Manuscript preparation was by M. Doughty, Dr. N. Eyles and Dr. C. Eyles.

Chapter 5 - Lake sediments as natural seismographs: earthquake-related deformations (seismites) in central Canadian (Ontario and Quebec) lakes

For this paper, all data collection in the field (seismic subbottom, sidescan and magnetometer data) was accomplished by M. Doughty (see following for exceptions). Dr. N. Eyles was in charge of logistics and navigation; additional field assistance was available for data collected on Lake Timiskaming, Lake Simcoe, Lake of Bays, Lake Joseph, Lake Muskoka and Lake Rousseau. Seismic data was collected for Lake Nipissing by Dr. N. Eyles (with field assistance). Seismic and sidescan data was collected for Lake Simcoe by Dr. N. Eyles, M. Doughty and Dr. J.I. Boyce. Magnetometer and bathymetric data was collected for Lake Simcoe by Dr. J.I. Boyce. Seismic, magnetometer and bathymetric data was collected for Lake Mazinaw by Dr. N. Eyles, M. Doughty and Dr. J.I. Boyce. Multibeam data was provided by the Canadian Hydrographic Survey (2005) for Lake Timiskaming. Extraneous datasets (e.g. remotely sensed imagery and digital elevation models) were assembled and manipulated by M. Doughty. All GIS work, seismic subbottom processing, sidescan imagery compilation and figure preparation was by M. Doughty (see following for exceptions). Figures 14B and 20 through 22 are originally by Dr. J.I. Boyce. Manuscript preparation was by M. Doughty, Dr. N. Eyles, Dr. C. Eyles and K. Wallace.

Chapter One

1.0 Introduction

Eastern North America experiences large intracratonic earthquakes that are not well understood but pose a risk to urban centers and other infrastructure (Fig. 1; Daneshfar and Benn, 2002; Kelson et al., 1996). These large earthquakes (M~7) are associated with fault movement along ancient rift systems that affect the integrity of the North American craton (Basham and Adams, 1989). One such rift system is the St. Lawrence Rift, first identified by Kumarapeli and Saull (1966) that extends from Lake Timiskaming in the north, through the Ottawa Valley and St. Lawrence River Valley with possible extension south-west into Lake Ontario and Lake Champlain (Figure 1). Unfortunately it is difficult to determine seismic hazard estimates for these rift systems, as large-magnitude events are not uniformly distributed, either spatially or temporally. Earthquake source zone models used to predict levels of seismic hazard for earthquake-resistant design of buildings, are based on a relatively short historical record, and tend to confine predicted future large earthquakes to zones of previously recorded large earthquakes (Basham and Adams, 1989). In areas where seismic potential could be underestimated through lack of historical seis-micity or where little is known about active faults, lake deposits can provide a valuable record of past seismic shaking events in the form of sediment deformation structures. Analysis of the form and distribution of sediment deformation structures may allow esti-mation of the magnitude and epicenter location of past earthquakes as well as identification of areas of potential future seismic risk (Karlin et al, 2004; Schnellman et al, 2002; Shilts et al, 1992; Doig, 1991).

Lake basins, across the Canadian Shield, are significant repositories of sediment in contrast to the scoured rock surfaces that surround them (Fig. 2). Surveys of such basins allow identification of glacial and Holocene depositional records that have not been preserved on exposed land surfaces (Eyles et al, 2003; Kaszycki, 1987). Exploring the stratigraphic infill of lakes in eastern North America has recently taken on increased importance with regard to identifying intracratonic seismic activity (e.g. Locat et al, 2003; Wallach et al, 1998; Shilts et al, 1989). Rift basins (grabens) can capture lengthy successions of Quaternary and Holocene sediment and may provide unbroken depositional records of considerable thickness. For example, the Kleszczow Graben in Poland contains an unbroken succession of between 300 and 400 m of Quaternary sediment (Brodzikowski et al, 1987), while the Baikal Basins within the Baikal Rift Zone of southeastern Siberia, contain a sedimentary infill between 4 and 7 km thick (Back and Strecker, 1998). Such lengthy and continuous depositional successions are rarely exposed but their characteristics and geometries can be investigated through application of sub-bottom/seismic reflection methodologies, multi-beam bathymetry and/or sidescan sonar technologies. Subaqueous and mass transport features such as slump and buried slump deposits, block slides, soft-sediment deformation features and megaturbidites can be identified in lake basin sediments and may be used as indicators of past seismic events (seismites; Karlin et al, 2004; Schnellmann et al, 2002; Shilts et al, 1992; Doig, 1991).

Landslides, sand blows and/or dykes (liquefaction features), fault offsets (horizontal; strike slip) and fault scarps (vertical; dip-slip) are possible surface expressions of seismic events (Sherrod et al, 2004; Locat et al, 2003; Stewart et al, 2001; Tuttle, 2001; Vanneste et al, 2001; Russ, 1982).

However, not all sediment deformation features described in late- and post-glacial lacustrine sediments can be ascribed to past seismic events as some may have resulted from other processes such as collapse due to karst phenomena, sediment compaction, dewatering and/or water escape, glaciotectonic thrusting below or at margins of ice sheets, melt of buried glacier ice or permafrost, and regional glacio-isostatic readjustments of the basin (Eyles et al, 2003; Brodzikowski et al, 1987; Shilts et al, 1992). During maximum glaciation, the upper crust was subjected to high compressional stress resulting in isostatic loading and the suppression of tectonic strain which in turn reduced the frequency of seismic events (Wu, 1998). Upon deglaciation, glacio-isostatic rebound enhances the potential for fault reactivation, likely causing enhanced seismic activity during the early, active crustal uplift phases (Adams and Basham, 1989) and decreased rebound stresses and seismic activity in the later post-glacial period. Faulting that affects both late-glacial deposits and extends through overlying modern deposits could be assigned a fault-related neotectonic origin (late-glacial or postglacial tectonic activity; Shilts and Clague, 1992) especially if these faults can be linked to displacements in the underlying bedrock. If such linkages with bedrock structure cannot be established, then it is unlikely that they could be distinguished from a glaciotectonic origin (Shilts and Clague, 1992). For surficial scarps and related features, certain diagnostic criteria have been suggested to discriminate neotectonic surface faulting from other features. Stewart et al. (2000) suggest that neotectonic surface faults can be recognized from 'offset of a surface or material that was originally continuous and unbroken, and whose dislocated fragments, if dated, can be demonstrated to be of equivalent age' (p. 1378, Stewart et al, 2000). Of primary importance for recognizing postglacial tectonic faults is their mechanism of formation and age. A postglacial fault should disturb or displace recent sediments or features (e.g. shorelines or streams) and have no evidence of glacial modification (Stewart et al, 2001).

Large underwater slumps (seen as chaotic to transparent seismic facies lacking continuous reflections), block slides and associated turbidite deposits recorded in thick successions of rift basin sediment may be reliable indicators of past major earthquakes. Underwater slumping events (with exceptions, based upon surrounding conditions such as initial slope angle) are more difficult to trigger than subaerial landslides as hydrostatic loading acts as a stabilizing force against slope failure (Karlin et al, 2004) requiring larger magnitude, or more proximal seismic events as the trigger. A widespread distribution of slump and related deposits was considered a key criterion for assigning an earthquake event as a trigger by Schnellmann et al (2002). Detection of these slumps and related deposits that result from strong seismic shaking using core and remote imagery such as high-resolution seismic profiling, multi-beam bathymetry and sidescan sonar, allows for the evaluation of the prehistoric seismic record for a region (Karlin et al, 2004; Locat et al, 2003; Schnellman et al, 2002; Vanneste et al, 2001; Shilts et al, 1992; Doig, 1991). Considerable strong seismic activity may have occurred during the early, rapid uplift phases of glacioisostatic rebound (Adams and Basham, 1989) and decreased rebound stresses and seismic activity in the later post-glacial period. Wu (1998) studied the causal relationship between intraplate earthquakes and postglacial rebound in eastern Canada and demonstrated that the spatial distribution of historical seismicity could not be explained by a strain rate distribution due to rebound alone. Both postglacial rebound and tectonic stress are required. A neotectonic origin, only, could be assigned to faulting that affects late-glacial or postglacial lake deposits and extends through overlying, recent sediments.

1.1 Objectives of this Research

This research aims to identify and analyze sediment deformation structures in lacustrine deposits within central Ontario and western Quebec in order to determine their mode of origin and the role of neotectonism, whether seismically or glacio-isostatically induced, in causing these structures to develop. The information presented here highlights the results of a multi-year seismic sub-bottom survey of lakes Gull, Muskoka, Joseph, Rousseau, Ontario, Wanapitei, Fairbanks, Vermilion, Nipissing, Lake Huron, Georgian Bay, Mazinaw, Simcoe, Timiskaming, Kipawa, Parry Sound and Lake of Bays, encompassing a total of more than 2000 kilometres of high-resolution track line data supplemented (in some cases) by multibeam and sidescan sonar survey records. Lake Temiskaming was the subject of detailed study given its proximity to seismically active zones along the St. Lawrence Rift System (Fig. 1).

Each of the individual chapters in this thesis contributes toward the overall research objective. Chapter 2 examines disturbed (slumped) lateglacial and postglacial lake sediments in Lake Kipawa resulting from the M6.2 1935 Timiskaming Earthquake. The mapping of lake floors for seismic risk analysis provides important data for the evaluation of 'actual' seismic risk in areas with short instrumental seismic records. Chapter 3 gives an introduction to research in the Timiskaming Graben (an extension of the Ottawa-Bonnechere Graben and part of the St. Lawrence Rift System) on the boundary of the Western Quebec Seismic Zone (Forsyth, 1981). The chapter provides a general overview of the research in the area (as an introduction to the more detailed study in the next chapter), and identifies the Timiskaming Graben as a 'weak zone' within the rigid North American craton (after Mazzotti, 2007). This chapter also provides tentative identification of active neotectonic deformation structures within the lake itself along with possible surficial features. Chapter 4 expands upon information first presented in Chapter 3 and provides detailed analysis of sediments and structures encountered within Lake Timiskaming. Chapter 5 extends the study into 15 other lake basins in central Ontario and, applying techniques demonstrated in the earlier chapters, examines and comments upon the concept of lake basins as 'natural seismographs' able to record postglacial and lateglacial sediment deformation. A broad-scale association between mid-plate seismicity and major suture zones and failed rifts is identified.

1.2 Thesis Structure

The objectives of this thesis are addressed within the second through fifth chapters. All chapters are formatted for publication in scientific journals and their contents are summarized immediately following.

Chapter 2 - Earthquake-triggered slumps (1935 Timiskaming M6.2) in Lake Kipawa, Western Quebec Seismic Zone, Canada

This paper (published in Sedimentary Geology; Doughty et al, 2010a) examines the effect of earthquakes on lateglacial and postglacial sediment infill of Lake Kipawa. This lake is located within the Western Quebec Seismic Zone (at the western boundary, within the Timiskaming Graben), a mid-plate region of eastern North America and an area of frequent, intracratonic earthquakes. The identification of widespread, large slumps in lake floor sediments indicate that the M6.2 1935 Timiskaming Earthquake was likely the largest in the area since deglaciation.

Chapter 3 - Ongoing neotectonic activity in the Timiskaming-Kipawa Area of Ontario and Quebec

This paper (published in Geoscience Canada; Doughty et al, 2010b), and meant for a non-specialist audience, describes the setting of the Timiskaming Graben (an extension of the Ottawa-Bonnechere Graben and St. Lawrence Rift System) and the preliminary results of the investigation of the lake floor sediments in Lake Timiskaming. Seismic data show two seismic-stratigraphic successions (identified as lateglacial and postglacial after Shilts, 1984) below the lake floor that are extensively deformed by horst and graben structures that control the lake topography/bathymetry. Deformation, affecting the modern lake floor, is ongoing and is interpreted to reflect seismic activity and associated crustal deformation.

Chapter 4 - Recent neotectonic faulting of lake floor sediments in an intracratonic rift and the implications for regional seismicity: Timiskaming Graben, Canada

This paper (published in Sedimentology; Doughty et al, 2013) expands upon the previous introductory paper of Chapter 3. Seismic reflection data and multi-beam bathymetry are used to describe features present upon and within the sediments of Lake Timiskaming and the likely processes responsible for their formation in order to differentiate between neotectonic and glaciotectonic structures. Crater-like sub-basins are interpreted to be 'glaciotectonic' resulting from the melt of trapped, large ice blocks and the subsequent collapse of overlying and adjacent sediment. Faulting, as seen on seismic section, is interpreted to be related to 'neotectonic deformation' associated with ongoing seismic activity.

Chapter 5 - Lake sediments as natural seismographs: earthquake-related deformations (seismites) in central Canadian (Ontario and Quebec) lakes produced by reactivation of Precambrian structures

This paper (in preparation, to be submitted to Tectonophysics) extends the analysis described in the previous chapter to 15 other lake basins in central Ontario. Here, seismic sub-bottom data collected from multiple lakes is examined (encompassing more than 2000 km of track lines) and used to supplement short seismic instrumental records. Various deformation structures (including, for example, faults, slumps, debris flows, water escape structures; collectively referred to as seismites) are interpreted to be of lateglacial age (truncated by a postglacial unconformity). This suggests the occurrence of moderate to large earthquakes during deglaciation (and rapid crustal rebound; e.g. Parry Sound, Lake Joseph and Lake Muskoka). Other features affect both lateglacial and postglacial sediments and appear to be related to ongoing seismicity due to crustal deformation (Lake Timiskaming and Lake Kipawa) or possibly related to activity along terrane boundaries (Lake Simcoe).

Chapter 6 – Conclusions and Recommendations for Future Work

This chapter summarizes the conclusions reached within this thesis and discusses potential directions for future research.

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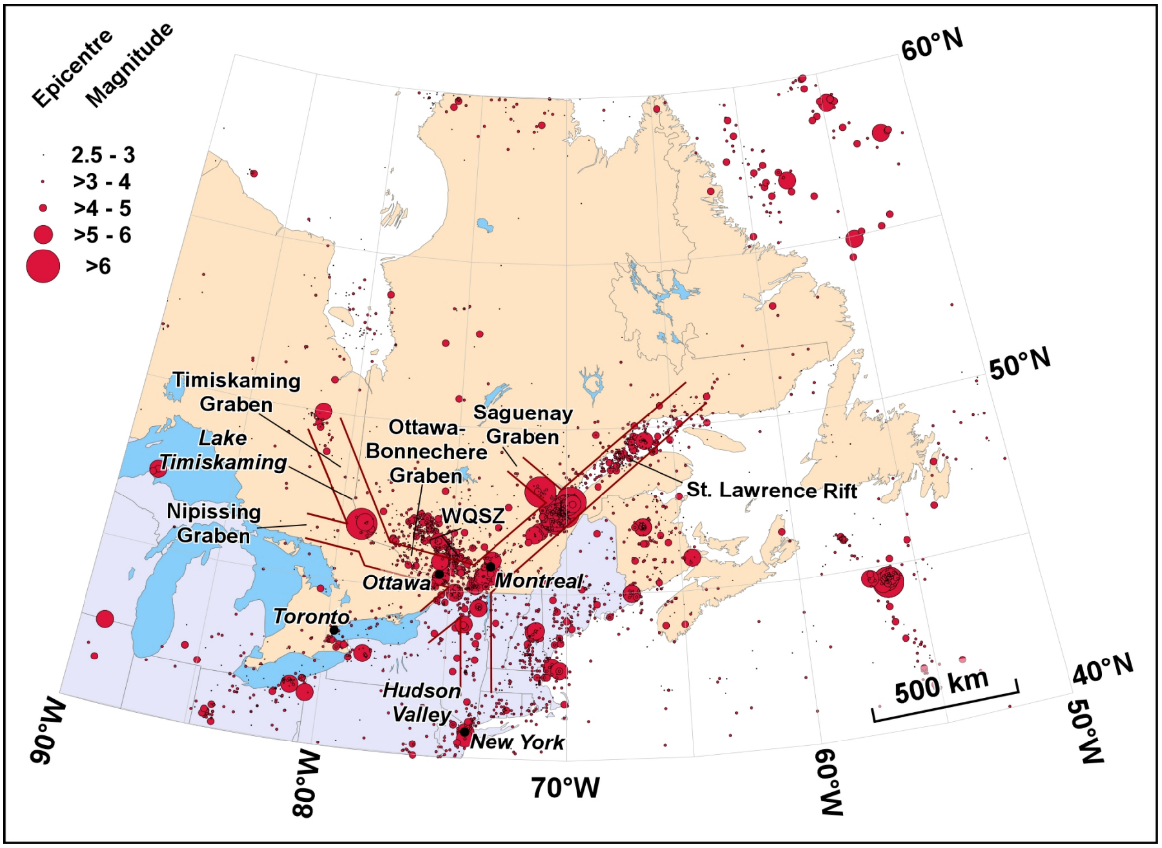

Figure 1: Earthquake epicentres in eastern North America (1627 to 2013 AD) and regional structural elements including the St. Lawrence Rift system and associated grabens (epicentres from Natural Resources Canada, 2013 and Halchuk, 2009). The Western Quebec Seismic Zone (Forsyth, 1981; WQSZ) is indicated.

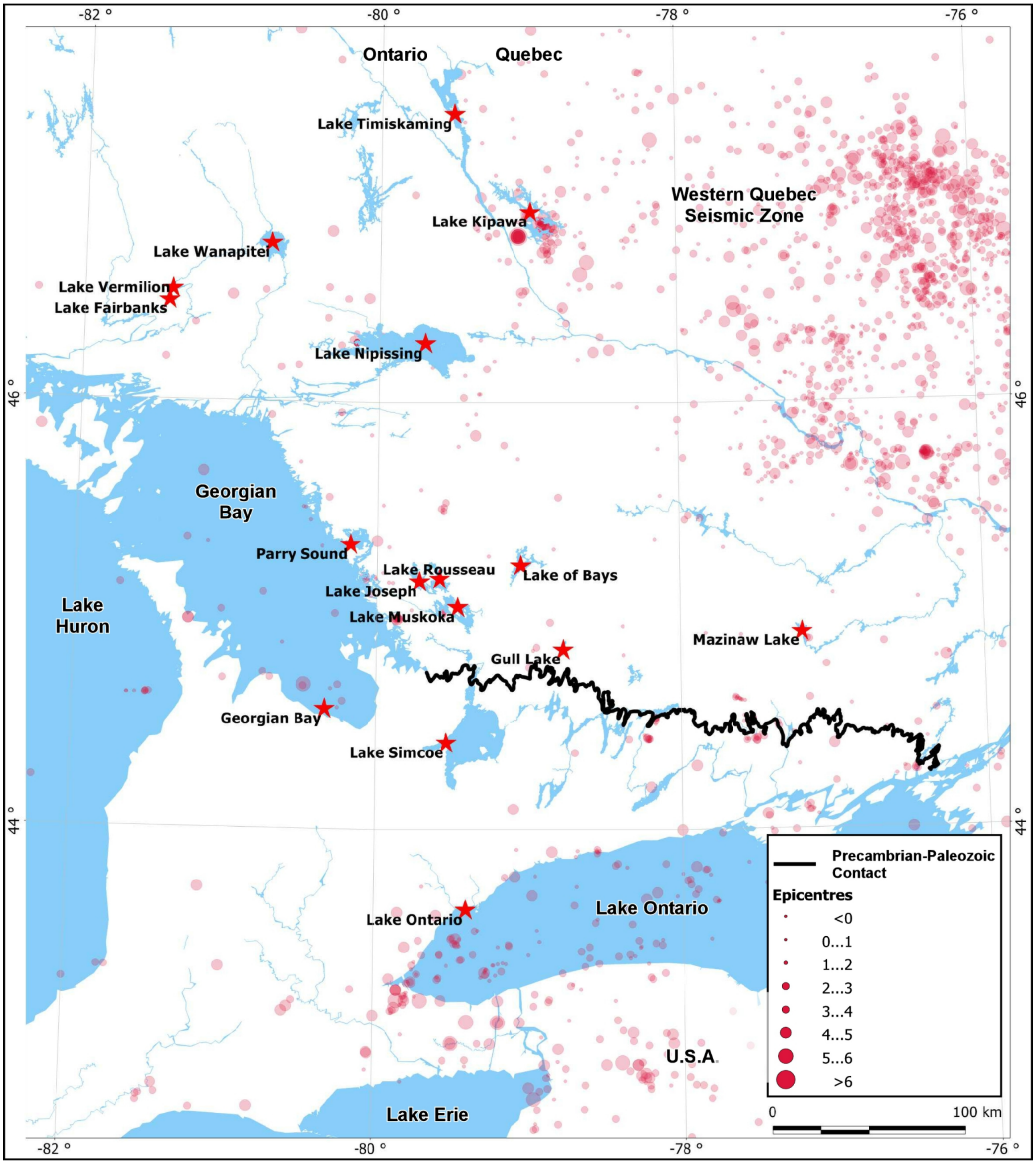

Figure 2: Study area. Data collection sites (red stars) are indicated along with earthquake epicentres (1627 to 2013; epicentres from Natural Resources Canada, 2013; Halchuk, 2009).

Chapter 2

Earthquake-triggered slumps (1935 Timiskaming M6.2) in Lake Kipawa, Western Quebec Seismic Zone, Canada

Doughty, M., Eyles, N. and Daurio, L. (2010) Earthquake (1935 Timiskaming M6.2) triggered slumps in Lake Kipawa, Western Quebec Seismic Zone, Canada. Sedimentary Geology, 228, 113-118.

ABSTRACT

The Western Quebec Seismic Zone (WQSZ) of eastern North America is characterised by frequent moderate magnitude intracratonic earthquakes (e.g., 1732, M5.8; 1935, M6.2 and 1944, M5.2). The WQSZ is centered along the Timiskaming and Ottawa-Bonnechere grabens, which form part of a complex aulacogen (St. Lawrence Rift) within the Canadian Shield. The WQSZ includes the urban areas of Montreal and Ottawa but seismic risk analysis is challenged by short instrumental records and long recurrence intervals. The M6.2 1935 Timiskaming Earthquake is the largest recorded to date and was felt over some 1.3 million km2 of eastern North America with many aftershocks of magnitude 4 to 5. Its epicenter lies below the western margin of Lake Kipawa, Quebec in the area where a major Proterozoic crustal boundary (the Grenville Front Tectonic Zone) crosses the Timiskaming Graben. A high-resolution ‘chirp’ seismic reflection survey of the lateglacial and postglacial sediment infill of Lake Kipawa reveals a clear record of recent ground shaking that is attributed to the 1935 earthquake. Widespread large slumps record down slope failure of the lateglacial and postglacial sediment fill indicating that the 1935 temblor was the largest in this area since deglaciation some 8000 years ago. Systematic mapping of landslides shows that they extend across an area of 600 km2 around the earthquake’s epicenter. Lakes cover a large area of eastern Canada; a regional-scale survey of lake floors could constrain historic epicenters and postglacial seismic history of the heavily populated WQSZ.

2.0 Introduction

Intracratonic earthquakes in the heavily populated areas of eastern Canada and the coterminous United States are a major cause for public concern and the focus of much research (e.g., Stein and Mazzotti, 2007). They are known to involve reactivation of Proterozoic structures such as failed rifts, now buried by younger Phanerozoic strata, as a consequence of ongoing plate movement and continued postglacial glacio-isostatic rebound (Kumarapeli, 1987; Wu, 1998; Mazzotti and Adams, 2005; Ma et al., 2008). Seismic risk analysis of intracratonic earthquakes is constrained by inadequate knowledge of seismic source zones, and lengthy recurrence intervals compared to the relatively brief duration of European settlement and short instrumental record (Basham and Adams, 1989; Johnson and Schweig, 1996; Stein and Mazzotti, 2007; Ma and Eaton, 2009). Such issues loom large across the Shield terrain of eastern Canada in connection with the safe design of mines and nuclear facilities including waste disposal sites. The stability of urban infrastructure on lateglacial marine sediments (Leda Clays) along the Ottawa and St. Lawrence Valleys is also a concern as these Leda Clays exhibit ‘quick’ behaviour when disturbed (Ma et al., 2008; Motazedian and Hunter, 2008; Rosset and Chouinard, 2008; Lundstrom et al., 2009). Attention focuses on the geologic and geomorphic record of seismic activity in eastern North America such as liquefaction structures, faults, sedimentary dikes and landslides (e.g., Kelson et al., 1996; Ouellet, 1997; Aylsworth et al., 2000; Talwani and Schaeffer, 2001; Tuttle, 2001). Lakes cover a significant portion (~ 12%) of eastern Canada and previous work has shown the utility of lakefloor mapping in identifying earthquake-related ground shaking and seismic risk (e.g., Shilts et al., 1989; Doig, 1991; Shilts and Clague, 1992; Shilts et al., 1992; Ouellet, 1997). In this regard, we present results of high-resolution seismic profiling of the floor of Lake Kipawa in the Western Quebec Seismic Zone where we identify a cluster of landslides associated with the epicenter of the 1935 Temiskaming Earthquake (M6.2).

2.1 Western Quebec Seismic Zone

The Western Quebec Seismic Zone (WQSZ; Forsyth, 1981) is a broad belt of enhanced seismic activity extending some 600 km from Lake Timiskaming southward along the Ottawa Valley of Canada as far as Lake Champlain in Vermont, USA (Fig. 1). It is associated with a large graben system in the North American craton (Canadian Shield) that includes the Timiskaming, Ottawa-Bonnechere grabens and their connection to extensional structures below the St.Lawrence Valley (the St. Lawrence Rift of Kumarapeli and Saull, 1966; Lovell and Caine, 1970) and the Hudson Valley of the USA extending to New York City (Fig. 1). The entire system is a complex failed rift structure resulting from one or more episodes of supercontinent breakup during the last 700 million years.

The Timiskaming Rift Valley straddles the Ontario/Quebec border forming a 60 km wide parallel-sided structure related to the Paleoproterozoic ‘Cobalt Embayment’ (a failed rift arm about 2400 Ma old) and has a long history of subsequent reactivation. Active when Laurentia broke free of Rodinia some 650 million years ago it was also affected by far field stresses from the Taconic Orogeny and other later collisional events beginning 440 million years ago (Dix and Molgat, 1998). The most recent extensional activity occurred after 200 Ma during the breakup of Pangea and the opening of the North Atlantic Ocean (Jackson and Fyon, 1992). Northwest-trending faults exert a strong structural control on regional topography and Paleozoic strata have been selectively preserved by subsidence along its northern reaches, having been stripped from the surrounding shield (Johnson et al., 1992).

The WQSZ includes Canada's national capital of Ottawa and the cities of Montreal and Cornwall and is one of the most active seismic areas in eastern North America with large damaging earthquakes in 1732 (Montreal; M5.8), in 1935 (Timiskaming; M6.2) and in 1944 (Cornwall–Massena; M5.6). The last was one of Canada's most damaging earthquakes with extensive impacts on urban infrastructure (Bent, 1996a). The epicenter of the 1935 Timiskaming Earthquake (the third largest on record in eastern Canada) is located 10 km east of Lake Timiskaming on the western margin of Lake Kipawa and was felt over some 1.3 million km^2 of eastern North America (Hodgson, 1936a, b; Bent, 1996b). Eight aftershocks of magnitude 4 to 5 occurred until March 1936. On average, M N 3 earthquakes occur every two years in the Timiskaming district; the most recent earthquake (M5.2) occurred on January 1, 2000 within 15 km of the 1935 epicenter (Adams et al., 2000). There is geomorphic evidence for recurring large (MN 6) Holocene earthquakes along the Ottawa Valley (Aylsworth et al., 2000). Doig (1991) reported near surface sediment layers disturbed by earthquake activity on the floor of Lake Timiskaming and Shilts

(1984) identified slumps produced by the 1935 Timiskaming Earthquake in Tee Lake just west of Lake Kipawa. Part of Lake Kipawa was briefly surveyed by Shilts (1984) who identified slumps from the 1935 event. The principal objective of our study was to systematically map the geographic extent of landslides in Lake Kipawa with reference to the recently constrained epicenter of the 1935 earthquake that is now placed along the western margin of the lake (Bent, 1996b) (Fig. 2).

2.2 **Physical Setting of Lake Kipawa**

Lake Kipawa occurs in an area of considerable crustal and structural complexity where the southeast trending Timiskaming Graben cuts across the Grenville Front Tectonic Zone (GFTZ) creating a ‘tectonic knot’. The Grenville Front is a major first-order crustal boundary of the North American craton separating the Archean Superior Province to the north from the Proterozoic Grenville Province and records the final assembly of Rodinia after c. 1.0 Ga when the Grenville was added to eastern North America (the ‘Grenville Orogeny’; see Tollo et al., 2004). Lake Kipawa fills a bedrock basin scoured by ice sheets into Proterozoic gneisses of the Grenville Province (Fig. 2). Physiographically, the area comprises typical Shield terrain with glacially-rounded gneiss bedrock knobs and many lakes. Lake Kipawa is some 65 km long north to south, with a total area of 300 km2 and more than 1000 km of rocky shoreline with many narrow (~300 m) intersecting sub-basins. The name Kipawa is derived from a Nishnabi aboriginal phrase ‘narrow passage between rocks’. Water depths reach a maximum of 98 m (Fig. 3) and the lake’s outlet (the Kipawa

River) discharges westward into Lake Timiskaming close to the GFTZ (Fig. 2). Lake Kipawa is a ‘black water’ Shield lake typical of eastern Canada lacking any major inflowing river. It is consequently starved of clastic sediment resulting in great water clarity and modern sedimentation dominated by gyttja; annual sedimentation rates are very low (~ 1 mm). The Kipawa area was flooded under the waters of a regionally extensive ice dammed lake (glacial Lake Barlow) during deglaciation some 9000 years ago (Dyke, 2004; Cronin et al., 2008).

2.3 Seismic Reflection Survey of Lake Kipawa and Identification of Slumps

We collected approximately 250 line kilometres of high-resolution seismic profiles across Lake Kipawa in 2007 using an EdgeTech X-STAR digital sub-bottom system employing chirp technology and a SB-216S tow vehicle (‘fish’). The system is ideally suited to investigations of lake floor deformations as it transmits a linearly swept ‘chirp’ pulse over a full frequency range of 2–12 kHz for 20 ms allowing for high-resolution mapping of bathymetry, profiling of bedrock topography, lake floor sediment stratigraphy and structure (e.g., Eyles et al., 2003; Lazorek et al., 2006); it is capable of resolving beds ~10 cm in thickness. Seismic reflection data below the flat lake floor shows two seismo-stratigraphic sediment successions. The oldest succession up to 30 m thick, shows closely-spaced high-frequency parallel reflectors and rests directly on glacially scoured bedrock filling the lower parts of sub-basins. This is typical of rhythmically-laminated glaciolacustrine silty-clays (‘varves’) deposited by underflows (quasi-continuous turbidity currents) in glacial Lake Barlow (Vincent and Hardy, 1979). The overlying sediment succession in Lake Kipawa is much thinner (~5 m) consisting of largely transparent, reflection-free organic-rich mud (gyttja) deposited primarily from suspension. This younger succession rests conformably on older sediments. The same twofold succession of lateglacial clastic rhythmites and postglacial gytjja is typical of most Shield lakes because they lack large modern river inputs and have minimal postglacial clastic sediment supply (Shilts and Clague, 1992; Lazorek, et al., 2006).

The seismic survey identifies many areas where the otherwise flat-lying sub-bottom stratigraphic couplet of late- and postglacial sediment is disturbed in the form of prominent mounds lying at the base of steep bedrock slopes (Fig. 4). These mounds are readily identified as the result of slumping (e.g., Lac Megantic - Shilts, 1984; Mulder and Cochonat, 1996; Halfman and Herrick 1998; Van Rensbergen et al., 1998; Locat and Mienert, 2003; ten Brink et al., 2006). Slumps extend between 100 and 300 m downslope from the base of steep slopes and are composed of largely transparent and reflection-free Holocene gyttja thickened by downslope landsliding during the 1935 earthquake (Shilts, 1984). Sidescan surveys proved to be of little value in mapping slump morphology given the transparent nature of postglacial sediment. Slumped sediments rest unconformably on underlying lateglacial sediment suggesting erosion during slumping (e.g., Lac Temiscouata; Shilts et al., 1992). This is suggested too, by the presence of blocks of Lake Barlow silty-clays within contorted postglacial sediment indicating that both lateglacial and postglacial sediments were involved in down slope failure. Fig. 3 shows the mapped distribution of slumps in Lake Kipawa including those identified in nearby Lac Tee by Shilts (1984).

2.4 **Discussion**

We cannot directly age date slumps in Lake Kipawa nor can we identify any differences in relative age within the total slump population. They are all of very recent age as they lie on the modern lake floor and lack any covering of sediment that is resolvable by high- resolution seismic mapping. Since we can resolve bed thicknesses to ~10 cm and given an annual sedimentation rate of ~1 mm, the uncertainty is 100 years or so. We are unable to identify any older slumps or debris flow facies preserved as hummocky deposits within the Holocene or lateglacial Barlow sediments of Lake Kipawa that might identify previous earthquake-related ground shaking events. These considerations suggest slumps are no older than 100 years and record an unusually large earthquake. Much current work on seismic risk analysis is concerned with hindcasting estimates of the intensity and magnitude of historic earthquakes from submarine landslides, such as the INQUA Earthquake Environmental Effects intensity scale (e.g., Michetti et al., 2004; Guerrieri et al., 2007). In the present case however, the well-documented 1935 Timiskaming Earthquake was the largest recorded in this area and its magnitude and effects are well known. It produced severe ground shaking across the entire Timiskaming district creating substantial structural damage to buildings and the collapse of railway embankments as far as 300 km distant (Hodgson, 1936a, b). More than 80% of the chimneys in the community of Témiscaming

10 km west of Lake Kipawa were damaged indicating a Modified Mercalli scale ranking of VII and severe ground shaking in the immediate vicinity of Lake Kipawa (Lamontagne and Bruneau, 1993). Local residents report rock falls from cliffs surrounding the lake and extensive discoloration of the waters of both Lake Tee and Lake Kipawa caused by resuspension of gyttja during widespread slumping of the lake floor (see Doig, 1991). The 1935 (M6.2) earthquake is one of the largest on record in the WQSZ with an inferred epicenter on the immediate western margin of Lake Kipawa along the southern boundary of the Grenville Front Tectonic Zone (Fig. 3) at a depth of 10 km. This lies within 15 km of the epicenter of a recent smaller (M5.2) earthquake in 2000 and occurs within a distinct cluster of located earthquakes since 1935 (Fig. 2). On average, the Timiskaming region experiences a M3 earthquake every other year (Adams et al., 2000). Although we are unable to verify this, it appears reasonable that slumping in lakes Tee and Kipawa is a product of the large 1935 earthquake (see also Shilts and Clague, 1992, p. 1031). Furthermore, as both glacial Lake Barlow and overlying postglacial sediments are included in slump deposits we conclude that this earthquake was the largest postglacial seismic event of the last 8000 years in the Kipawa area.

Lastly, we note a preferred structural association between the epicenter of the Timiskaming Earthquake and the intersection of the Timiskaming Graben with the Grenville Front Tectonic Zone (Fig. 3). Slumps occur preferentially in that part of the Lake Kipawa basin south of the Front that is cut into Proterozoic Grenville gneisses. These gneisses have a markedly ‘whaleback’ topography of mammilated bedrock highs and steep side slopes. This geographic association might indicate a distinct structural, lithological or even geomorphological control on the effects of shaking, which we wish to explore by expanding this study to surrounding lakes. One objective of such work will be to establish any systematic variation in the size of slumps across lake basins (e.g., ten Brink et al., 2006).

2.5 **Summary and Conclusions**

The Western Quebec Seismic Zone is one of the most seismically active areas in eastern Canada and adjacent areas of the USA with several large cities at risk. Seismicity in this part of the Canadian Shield is the product of the reactivation of a complex aulacogen structure within the St. Lawrence Rift System that crosses the grain of the Grenville Front Tectonic Zone. Risk analysis is retarded by a lack of information regarding the occurrence and frequency of postglacial earthquake activity because of short instrumental records. Recognition of earthquake-related slumps on lake floors and the importance of mapping lake floors for seismic risk analysis are emphasized by Shilts and Clague (1992) and Ouellet (1997) as a tool for constraining seismic risk. Our study of Lake Kipawa identifies a large number of slumps on the lake floor located around the known epicenter of the 1935 Timiskaming Earthquake (M6.2) south of the Grenville Tectonic Front where the Superior Province abuts the Grenville Province. Slumping has affected both deposits of glacial Lake Barlow (dated at c. 8000 years before present) and an overlying postglacial drape of gyttja suggesting that the 1935 earthquake was the largest postglacial event in this area. Comprehensive high-resolution seismic mapping of the many lake basins in eastern Ontario and Quebec could constrain the epicentral location and recurrence interval of large postglacial earthquakes across the Western Quebec Seismic Zone.

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List of Figures

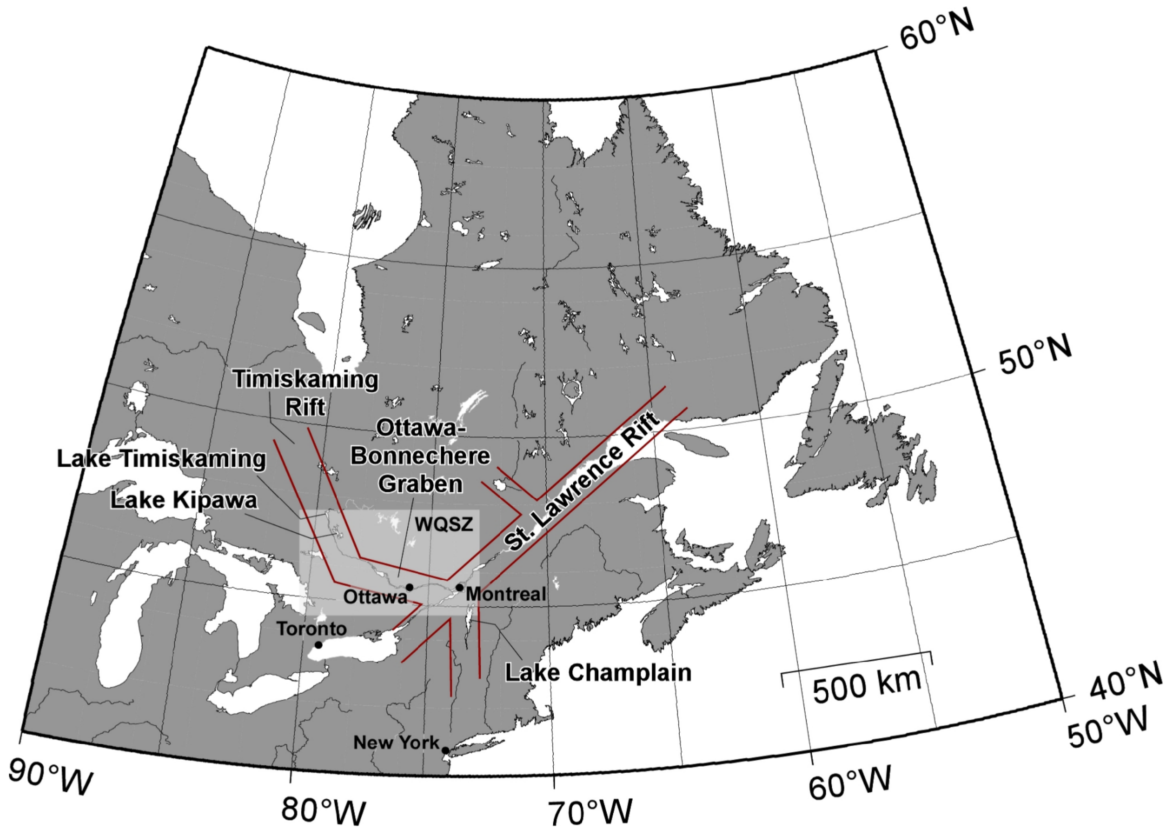

Figure 1. Regional structural setting of the Western Quebec Seismic Zone and St.Lawrence Rift System in eastern North America.

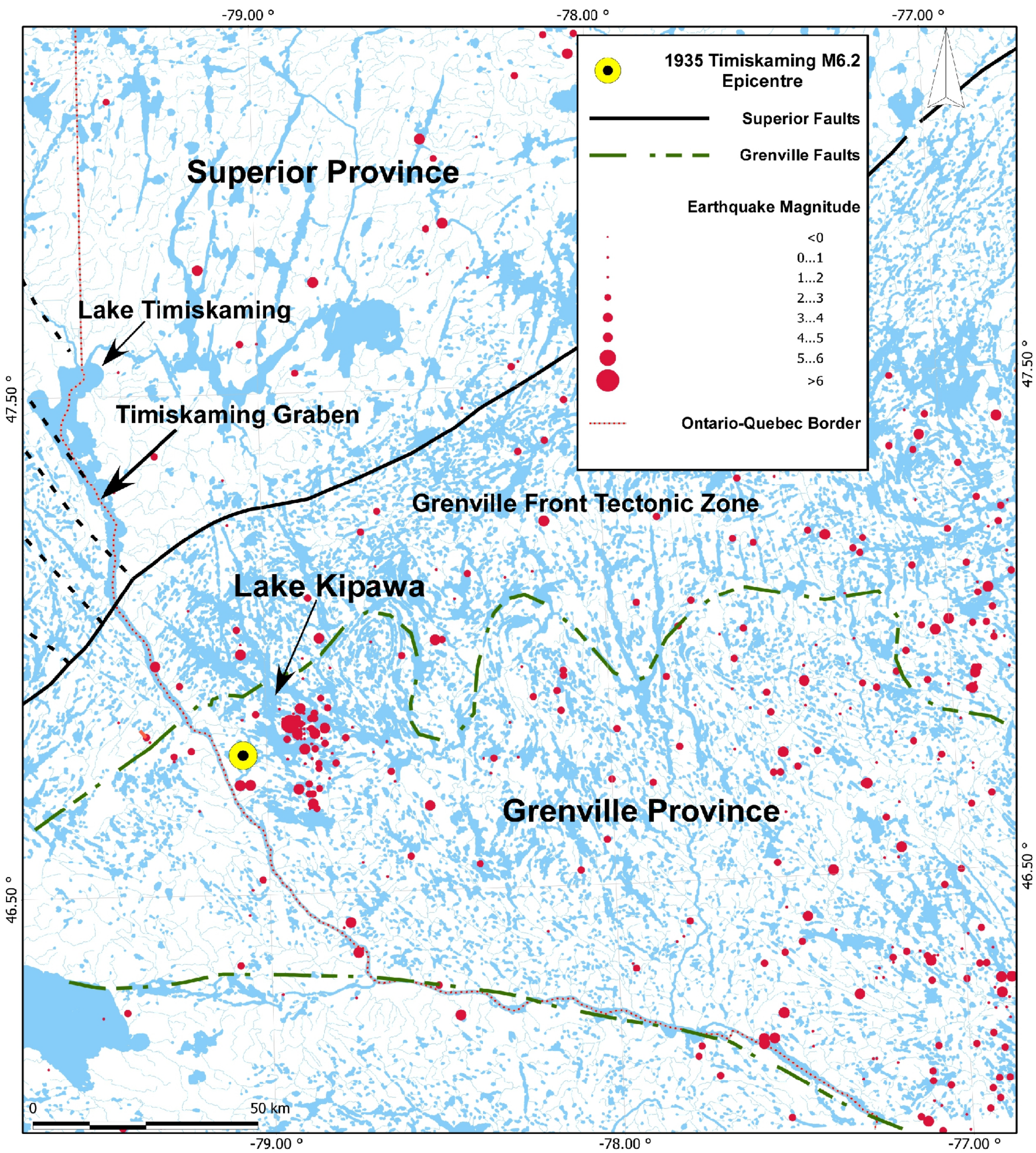

Figure 2. Regional map of Lake Kipawa, Quebec showing recorded epicenters of seismic events and the epicenter of the 1935 M6.2 Timiskaming Earthquake. Superior and Grenville Province faults are also shown.

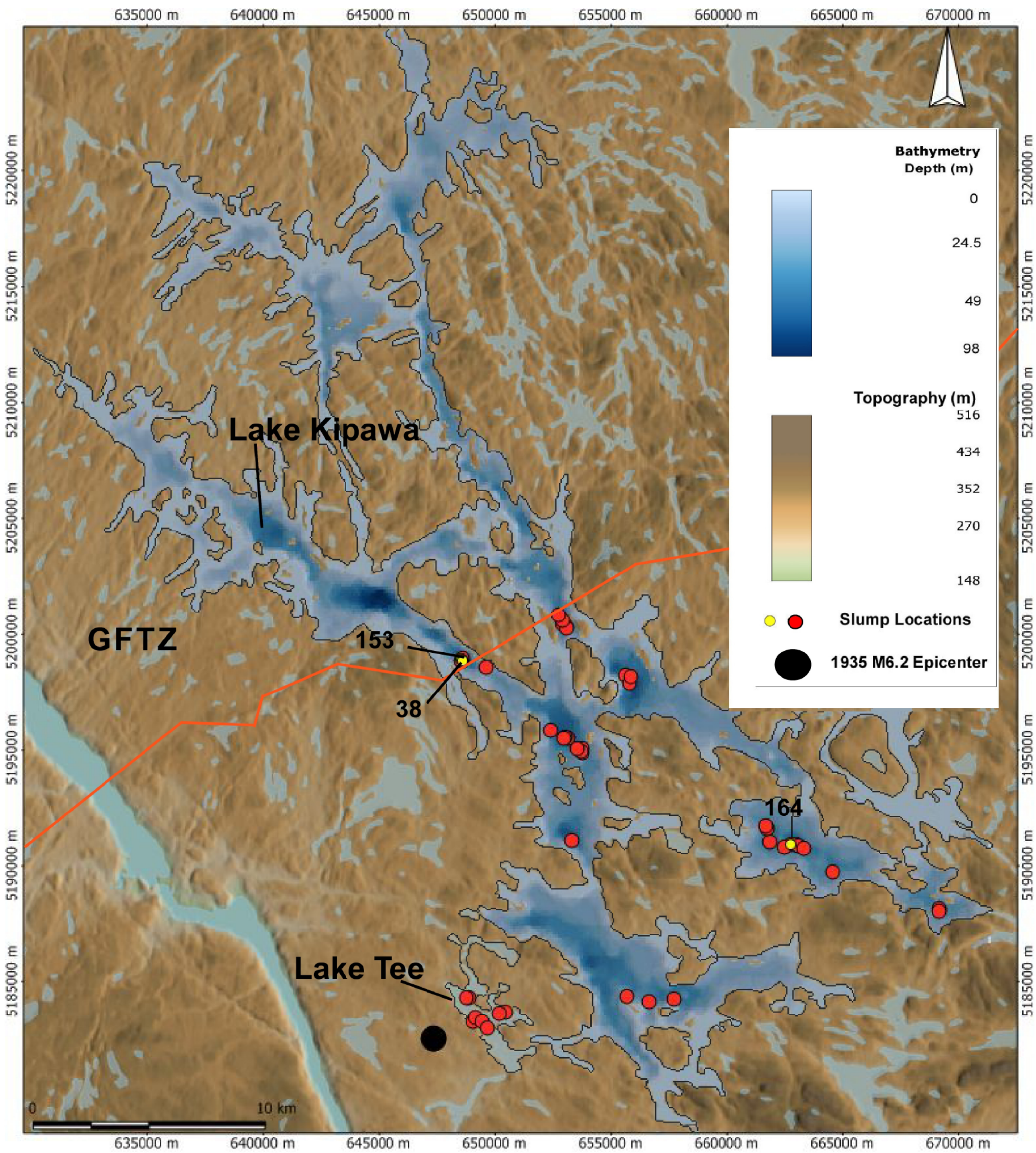

Figure 3. Bathymetry of Lake Kipawa, Quebec showing epicenter of 1935 Timiskaming Earthquake and locations of mapped slumps on lake floor. Position of Grenville Front Tectonic Zone dividing the Superior Province from the Grenville Province is also shown. Bathymetry from Quebec Ministere des Ressources Naturelles (2000). Digital elevation data from Jarvis et al (2008). Lac Tee slumps after Shilts (1984).

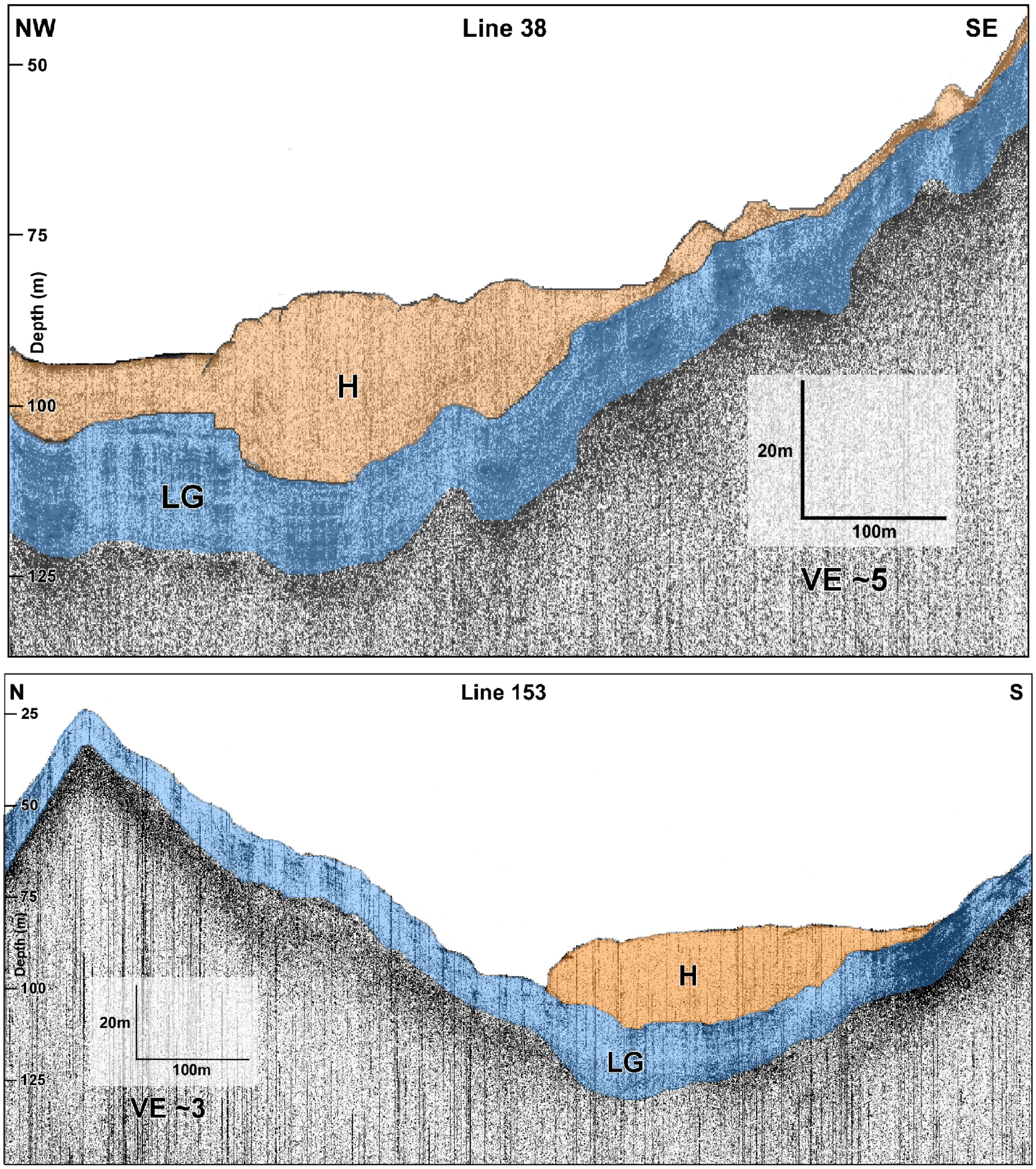

Figure 4. High-resolution seismic profiles from Lake Kipawa of slumped Holocene (H) ‘gyttja’and lateglacial (L/G) silty-clays at sites. Sediment and water depths assume a constant water column velocity of 1550 m/s and average sub-bottom velocity of 1650 m/s (e.g., Mullins and Eyles, 1996).

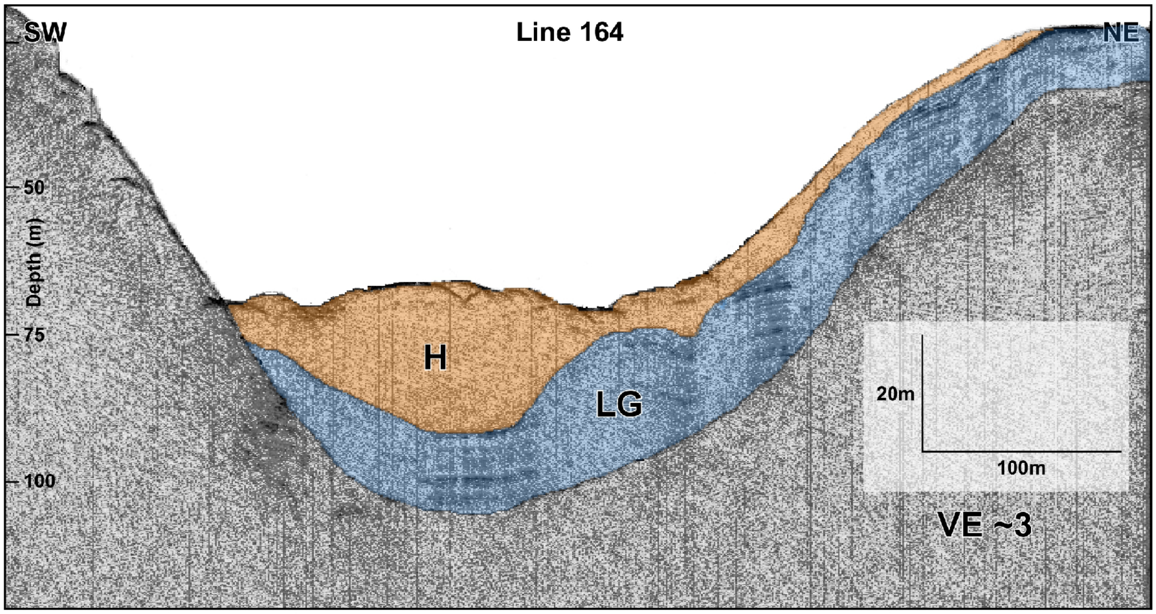

Figure 4. Continued ...

Chapter 3

Ongoing neotectonic activity in the Timiskaming-Kipaawa Area of Ontario and Quebec

Doughty, M., Eyles, N. and Daurio, L. (2010) Ongoing neotectonic activity in the Timiskaming-Kipawa area of Ontario and Quebec. Geoscience Canada, v 37, p. 97-104.

SUMMARY

The Timiskaming Graben lies along the border of Ontario and Quebec within the Western Quebec Seismic Zone a conspicuous belt of heightened intracratonic seismic activity in eastern Canada. The graben forms a conspicuous 50 km wide fault-bounded morphotectonic depression partly filled by Lake Timiskaming (ca. 100 km long, 209 m max depth). This lake is a postglacial successor to much larger glacial Lake Barlow, which drained about 8000 years ago leaving an extensive clay plain ('Little Clay Plain'). Some 1000 kilometres of high resolution seismic sub-bottom data collected from Lake Timiskaming reveal that both lateglacial Barlow and Holocene sediments are extensively deformed by neotectonic horst and graben structures. The bathymetry of the lake floor is structurally controlled by faulting; graben basins record enhanced postglacial subsidence between parallel bounding faults, one of which is expressed onshore on the surrounding clay plain as a 20 km-long, 10 m high scarp. These structures indicate ongoing neotectonic activity on a scale not recognized elsewhere across intracratonic North America. Seismic reflection data confirm the Timiskaming Graben as an intraplate 'weak zone' that may contain a long Late Cenozoic sediment record. A program of deep continental drilling within the Timiskaming Graben and extension of the current program of investigating lake-floor geology across the many lakes of the Western Quebec Seismic Zone are now needed.

3.0 Introduction

The M 5 Earthquake of June 23rd, 2010 centred 60 km northeast of Ottawa within the Western Québec Seismic Zone (Fig. 1), rattled much of central Canada and is a reminder that intracratonic seismicity in heavily urbanized central and eastern North America is still not well understood (Stein and Mazzotti 2007). The Western Québec Seismic Zone poses a direct risk to infrastructure such as mines, lifelines, critical facilities and the large urban areas of Montréal and Ottawa, where soft glaciomarine sediments (Leda Clay) and aged infrastructure compound seismic risk (Motazedian and Hunter 2008; Rosset and Chouinard 2008). There is still inadequate knowledge of earthquake source zones, recurrence intervals, and the effects of postglacial rebound (Ma et al. 2008), so emphasis is placed on the geologic and geomorphic record of seismic activity (e.g. Shilts and Clague 1992; Kelson et al. 1996; Ouellet 1997; Aylsworth et al. 2000; Tuttle 2001; Talwani and Schaeffer 2001). Our purpose here is to highlight the wealth of neotectonic structures that deform the bottom sediments of lakes Kipawa and Timiskaming[1] within the Timiskaming Graben along the border of Ontario and Québec (Fig.2). We also identify a prominent 20 km-long onshore scarp that is inferred to be a product of co-seismic faulting.

[1] Timiskaming is variably spelled Temiscamingue, Temiskaming and Temiscaming in Ontario and Quebec. For simplicity we use Timiskaming throughout.

3.1 Timiskaming Graben

The Timiskaming Graben (TG) parallels the Ontario–Québec border and is structurally defined by the Cross Lake Fault and the Quinze Dam Fault (Fig. 2). It is a complex structure that likely first developed as part of the Paleo-proterozoic (ca. 2.4 Ga) 'Cobalt Embayment'; it was much later reactivated during regional extension related to the break-up of Rodinia and Pangea (Barosh 1990; Mazzotti 2007) and during compression related to various Appalachian orogenies (Kumarapeli and Saul 1966; Dix and Molgat 1998). The TG transects the Grenville Front Tectonic Zone, which separates the Grenville and Superior provinces, and merges with the Ottawa-Bonnechere Graben before joining the St. Lawrence Rift (Fig. 2). The entire system is a broad belt of heightened seismic activity known as the Western Québec Seismic Zone (Fig. 1), which encompasses the large urban areas of Ottawa–Hull and Montréal, and extends to Lake Champlain and Vermont, USA (Fig. 2). The history of earthquake activity in this area is therefore of international concern.

Basement to TG consists of Paleoproterozoic and Archean rocks of the Canadian Shield (Lovell and Caine 1970; Lovell and Frey 1976; Russell 1984; Johnson et al. 1992; Dix and Molgat 1998). The graben preserves outliers of Ordovician to Silurian limestones of the Upper Ordovician Liskeard Group and Lower Silurian Wabi Group (Dix et al. 2007) that are downfaulted by as much as 300 m into older Shield strata (Fig. 2). Glacial Lake Barlow was located within the confines of TG between 9000 and 8000 years before present (BP), during Laurentide Ice Sheet deglaciation (Barnett 1992; Dyke 2004; Cronin et al. 2008). The exposed floor of glacial Lake Barlow is referred to as the ‘Little Clay plain’ (informal term; Fig. 2) and is underlain by glaciolacustrine silty clays (varves). Lake Timiskaming (100 km long, 200 m maximum depth) is the postglacial successor to glacial Lake Barlow; Barlow laminated silt-clays are present below the floor of Lake Timiskaming as far south as the McConnell Moraine (Fig. 2). The moraine was deposited on a topographic ridge along the Grenville Front Tectonic Zone and forms the southern termination of glacial Lake Barlow (Veillette 1994).

3.2 **High Resolution Seismic Images of Neotectonic Structures in Lake Timiskaming**

One thousand kilometres of high-resolution seismic data were collected on Lake Timiskaming using an EdgeTech X-STAR digital seismic profiling system (see Lazorek et al. 2006 for details). Multibeam data collected by the Canadian Hydrographic Survey (Canadian Hydrographic Survey 2004) were also used to relate bathymetry to deformation structures seen on seismic profiles. Lake Timiskaming is a large lake - 14 km wide in its northern part, and some 100 km in length. To the north, its linear western margin is sharply definside by shallow (<5 m) wave-cut ‘shelves’ and several limestone islands and intervening bays (Fig. 3).

Seismic data reveal two seismic-stratigraphic successions below the floor of the lake (Fig. 5). Parallel high-frequency reflectors typical of varved glacial Lake Barlow silty clays characterize the older succession (Shilts 1984), which is overlain by a largely transparent, reflector-free Holocene succession. Within Lake Timiskaming, the younger succession is unusually thick (40 m) compared to other glaciated ‘Shield’ lake basins (Shilts and Clague 1992; Lazorek et al. 2006) because of high sediment inputs from several large inflowing muddy rivers that rework glacial Lake Barlow sediment. The seismic survey shows that the late-glacial and postglacial sediment fill of Lake Timiskaming is extensively and severely deformed by horst and graben structures that control the topography of the lake floor (Figs. 4, 5A-E). The remarkably linear, parallel-sided, central deep-water basin of Lake Timiskaming (Fig. 3) is a trench-shaped graben resulting from subsidence between the Timiskaming West Shore Fault (TWSF) and another hitherto undiscovered fault that is named herein the Timiskaming East Shore Fault (Figs 2, 3, 4). The latter trends north–south, parallel to the TWSF through Dawson Point and the west side of Mann Island and actually consists of a zone of closely spaced faults (Fig. 5B, C). Shallow-water bays on the eastern margin of the lake, such as Baie Joanne (Fig. 3) and Baie Paulson (Fig. 4A) are the expression of small grabens containing downfaulted late-glacial and postglacial sediments, lying between limestone islands that form horsts (Fig. 6). A detailed 3-D reconstruction of faults is being presented elsewhere, but we can report sidescan records of several examples of open lake-floor fissures above faults, indicating recent or ongoing activity (Fig. 4B, C). Deformation is not restricted to a short phase of heightened glacioisostatic rebound immediately following deglaciation, as reported elsewhere in intracratonic settings (e.g. Lagerback 1992), because such structures would be buried below modern sediment. Instead, deformation has affected the modern lake floor and is ongoing.

Faulting in the northern part of Lake Timiskaming was briefly reported by Shilts (1984) and attributed to late-glacial glaciotectonic processes involving collapse over buried ice blocks, although it was noted that postglacial sediments had also been displaced. Shilts and Clague (1992) later suggested a link to the M 6.2 Timiskaming Earthquake in 1935. Doig (1991) reported deformation structures in short drill cores and attributed the deformation to seismic shaking. Our substantial seismic coverage demonstrates a pronounced structural control on modern lake-floor topography that is consistent with earthquake-related neotectonic activity. In particular, the faulting of both postglacial and late-glacial sediments indicates a neotectonic, not glaciotectonic origin.

Tectono-geomorphic features on the surface of the Little Clay plain appear to provide evidence for rapid subsidence during at least one postglacial earthquake. A remarkably linear, 20 km long, 10 m high, northwest-trending, west-facing scarp between Thornloe and New Liskeard (Fig. 7) coincides with the northern extension of the Timiskaming East Shore Fault interpreted from offshore seismic profiles. Previous workers had suggested that the scarp is the surface expression of a bedrock fault (Lovell and Frey, 1976; Russell 1984) whereas others mapped it as an abandoned glacial Lake Barlow shoreline bluff (Morton et al. 1979). Beach deposits were not observed along the length of the scarp, whose striking linearity is markedly different from other shorelines on the Little Clay plain. The onshore scarp and the Timiskaming East Shore Fault likely record one or more large prehistoric earthquakes and subsidence of the central basin between the

Timiskaming West and East Shore faults (Fig. 3). Subsidence of the Little Clay plain has promoted the growth of extensive peat deposits, some of which are covered by 'run-out debris' from mudflows that record local landsliding of glaciolacustrine silty clays down the scarp face. It is possible that these mudflows may have neotectonic significance; i.e. they may have been triggered by earthquake activity. Possible 'sand blows' or sand volcanoes (Tuttle 2001) where sand has been injected upward through peat, are noted (Fig. 8). Examination and coring of the buried peat stratigraphy below landslide deposits, and in sand blows, may ultimately yield a dateable record of tectono-geomorphic activity.

Evidence of earthquake-triggered sediment slumping that can be related to a specific earthquake is found on the floor of Lake Kipawa, east of Lake Timiskaming (Doughty et al. 2010; Figs 1, 9). Like Lake Timiskaming, its sediment infill consists of Barlow silty clays (maximum thickness 30 m) overlain by a thin (<5 m) succession of organic-rich Holocene sediment. High-resolution seismic profiling on Lake Kipawa (not shown) reveals some 35 distinct locations where the sediments of both successions have experienced downslope failure; these large slumps are characterized by hummocky mounds of thickened and contorted deposits. Shilts (1984) reported their presence in Lake Kipawa and also recognized them in nearby Lake Tee (Fig. 9). Detailed mapping of these landslides reveals that they occur within a broad 600 km2 zone around the epicentre of the 1935 Timiskaming Earthquake (Fig. 9).

3.3 **Wider Significance**

To our knowledge, the scale and type of active neotectonic deformation in the Timiskaming Graben briefly reported herein is, so far, fairly unique in eastern North America (see Adams et al., 1991, for exceptions; Kelson et al. 1996; Tuttle 2001; Talwani and Schaeffer 2001; Stein and Mazzotti 2007). The deformation is interpreted to reflect frequent seismic activity and associated crustal deformation in the Western Québec Seismic Zone. This finding is consistent with Mazzotti's (2007) identification of the Timiskaming Graben as a persistent 'weak zone' within the otherwise rigid North American craton. Kumarapeli and Saul (1966) compare the Proterozoic–Phanerozoic development of the St. Lawrence Rift to the modern East African Rift, and, indeed, the graben infill of Lake Timiskaming is similar to that in other rift basins worldwide (e.g. Scholz et al. 1998; Gawthorpe and Leeder 2000; Withjack et al. 2002).

The Western Québec Seismic Zone is one of the most active seismic zones in eastern North America; large, damaging earthquakes occurred in 1732 (Montreal; M 5.8), in 1935 (Temiscaming; M 6.2) and in 1944 (Cornwall; M 5.6). The epicentre of the 1935 earthquake was located 10 km east of Lake Timiskaming near Lake Kipawa and is recorded by extensive lake floor slumping (Hodgson 1936; Bent 1996; see previous section). The 20-km-long scarp between Thornloe and New Liskeard appears to mark subsidence of the surface of the Little Clay plain along the Timiskaming East Shore Fault (Fig. 7), and points to a heightened seismic risk for the communities of New Liskeard and

Haileybury, which lie on the same downfaulted block but just above lake level (Fig. 6). Earthquakes (M>3) occur every two years in the Timiskaming district (Adams et al. 2000). Geomorphic evidence for recurring, large (M>6) Holocene earthquakes along the Ottawa-Bonnechere Graben (Fig. 2) was identified by Aylsworth et al. (2000) and is consistent with the data reported here from the Timiskaming Graben. The most recent earthquake (M 5.2) in the Timiskaming area occurred on January 1, 2000 under Lake Kipawa, within 15 km of the 1935 epicentre. This scale of activity is not consistent with a model of seismicity driven by postglacial rebound alone (e.g. Chung 2002) but is more consistent with ongoing deformation within the North American Plate as a consequence of the interaction of plate movement and old structures (Mazzotti 2007).

The flat floor of glacial Lake Barlow (i.e. the Little Clay plain) is an ideal substrate for the rapid identification of neotectonic structures and fault offsets (e.g. Hume 1925) in a way not possible on the rough exposed bedrock of the surrounding Shield surface. It is apparent that a wide range of structurally controlled geomorphic features occur in the area and await detailed investigation (Figs. 4, 5). Given the implications of the findings presented herein for hazard assessment of the cities of Ottawa and Montréal (Motazedian and Hunter 2008; Rosset and Chouinard 2008), a program of tectono-geomorphic mapping is being conducted across the Timiskaming Graben and Little Clay plain. Sub-bottom seismic reflection profiling will also be carried out on the many other lakes in the region, to constrain the record of postglacial seismicity in the Western Québec Seismic Zone. The Timiskaming Graben is an ideal candidate for deep continental drilling.

It is possible that a long, late Cenozoic sediment and possible paleoclimate record is preserved below the central graben basin of Lake Timiskaming. Deep drilling would also constrain the long-term tectonic history of the graben.

Acknowledgements

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List of Figures

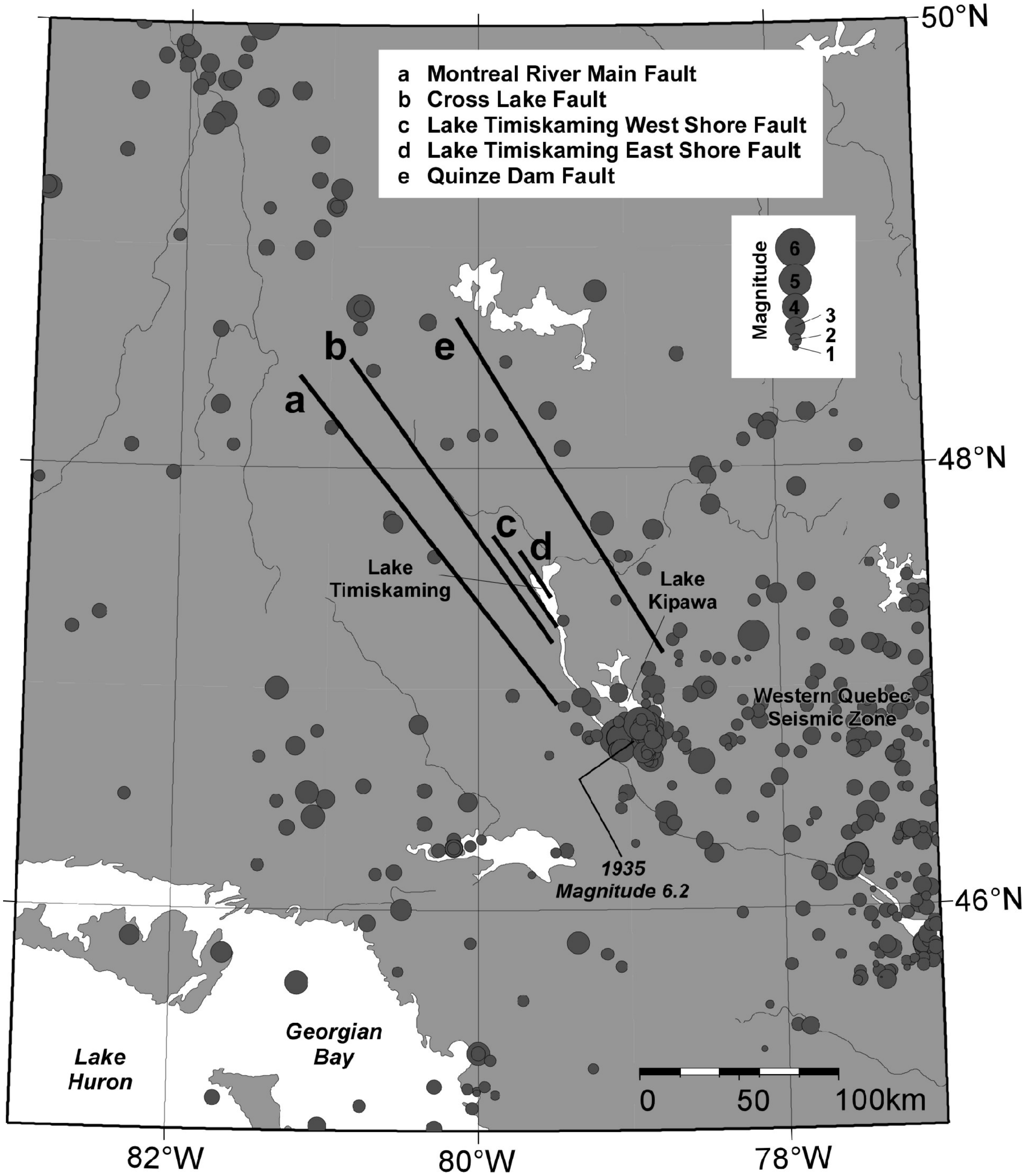


Figure 1. Earthquake epicentres within the Western Québec Seismic Zone (source: Geological Survey of Canada) and principal faults (a to e) associated with Timiskaming Graben (Fig. 2).

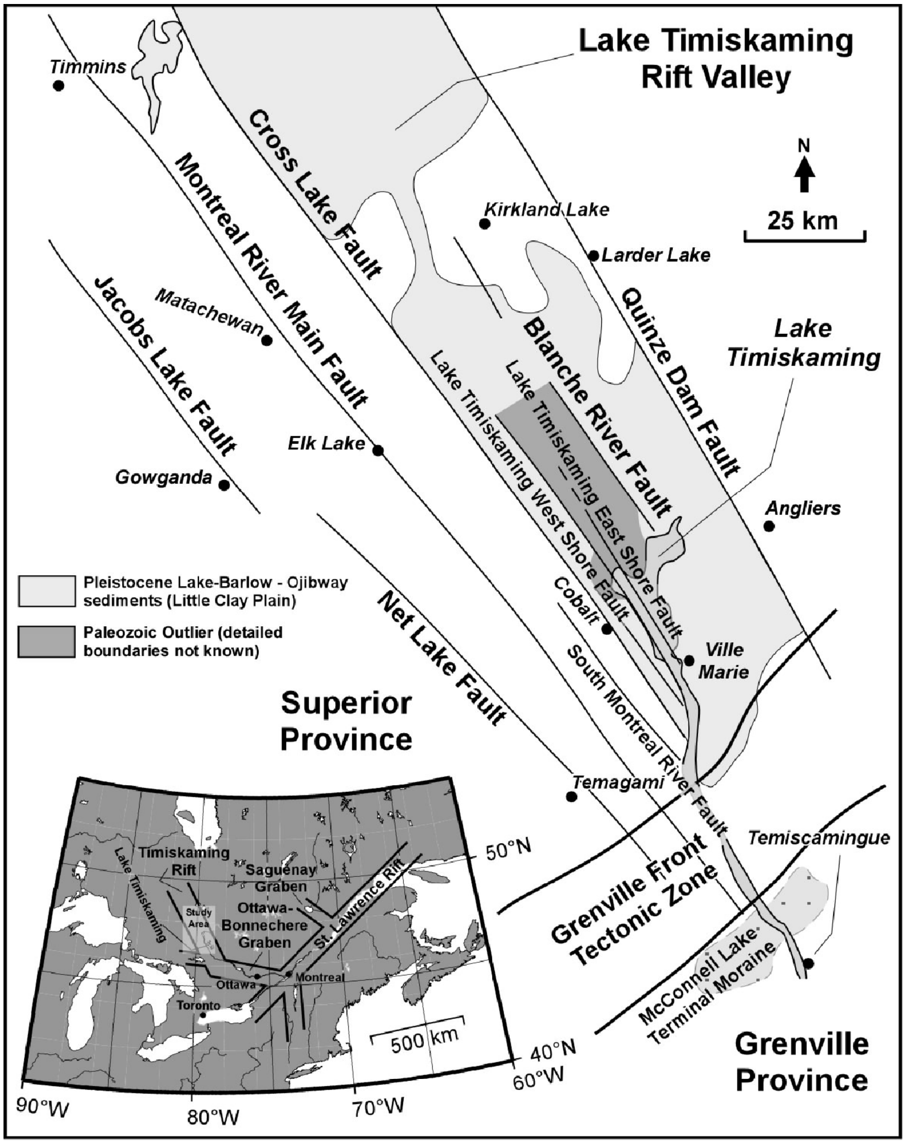


Figure 2. Simplified geological and physical setting of Timiskaming Graben along the Ontario–Québec border. A large block of early Paleozoic limestone (Upper Ordovician Liskeard Group and Lower Silurian Wabi Group; Dix et al. 2007) is downfaulted into the Superior Province between the Timiskaming West Shore Fault and the Blanche River Fault. The deep central basin of Lake Timiskaming is controlled by the Timiskaming West and East Shore faults (Fig. 3).

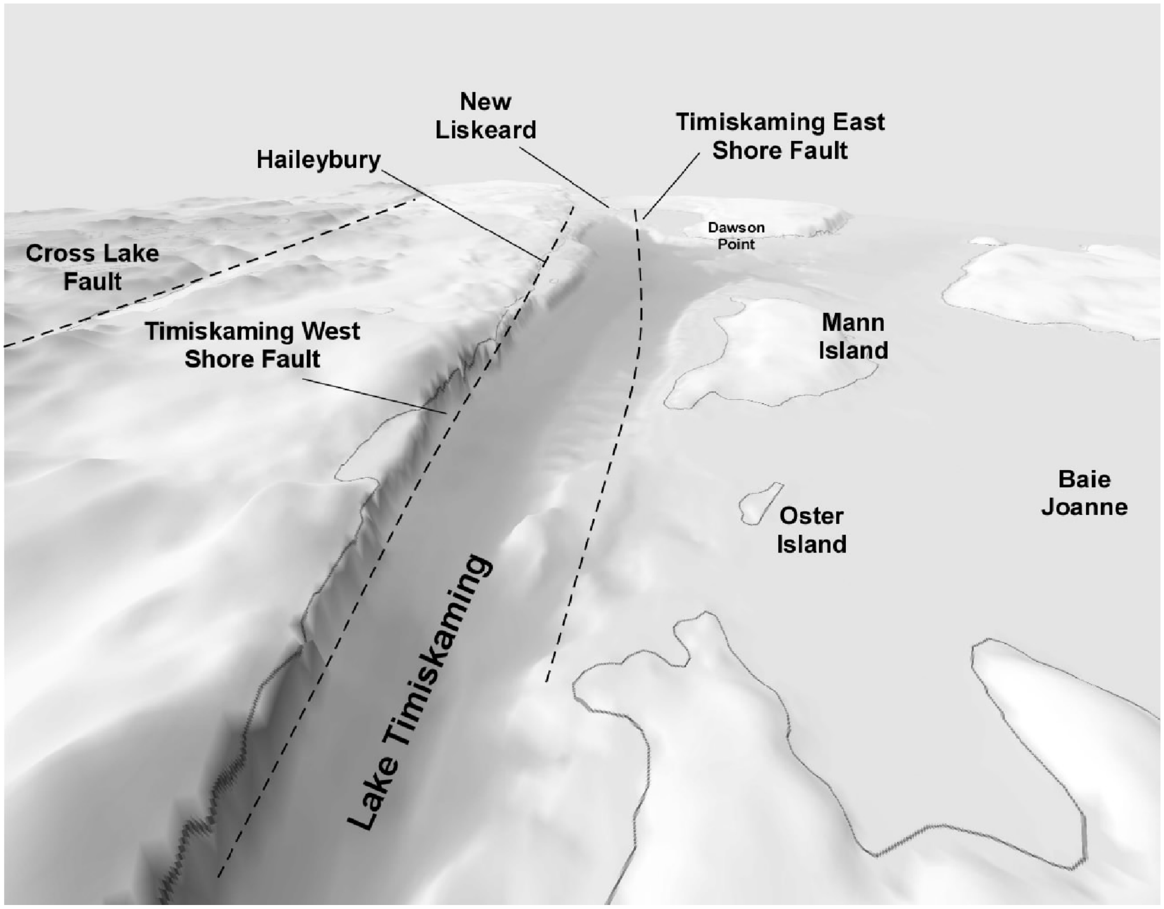


Figure 3. Perspective view of the bathymetry of the northern part of Lake Timiskaming, looking north from point X on Figure 4A (data provided by Canadian Hydrographic Service). A deep central graben basin is defined by the Timiskaming West and East Shore faults.

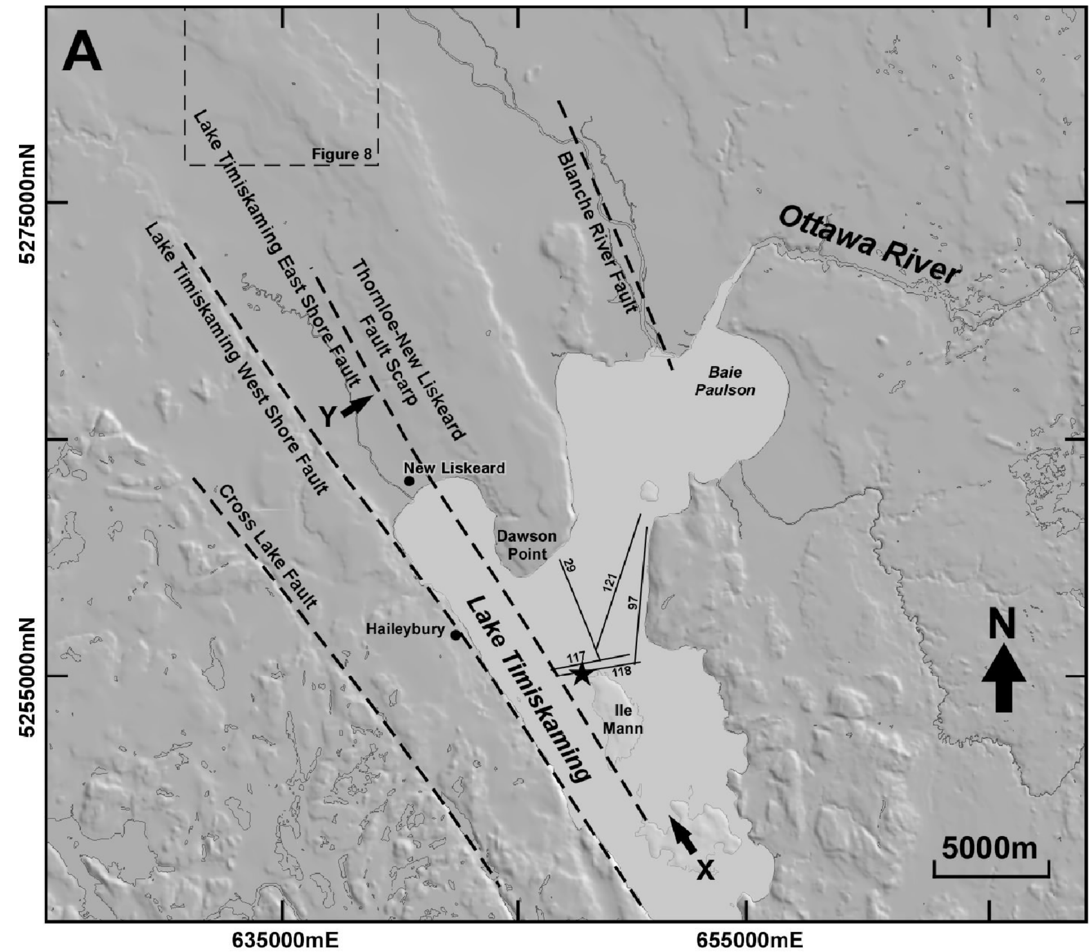


Figure 4. A. Shaded-relief image of northern end of Lake Timiskaming showing locations of offshore seismic track lines (Fig. 5A–E) and the Thornloe–New Liskeard fault scarp on the east side of the northerly extension of the Timiskaming East Shore Fault. Points X and Y and prominent arrows indicate viewpoints and 'look' directions for Figures 3 and 7, respectively.

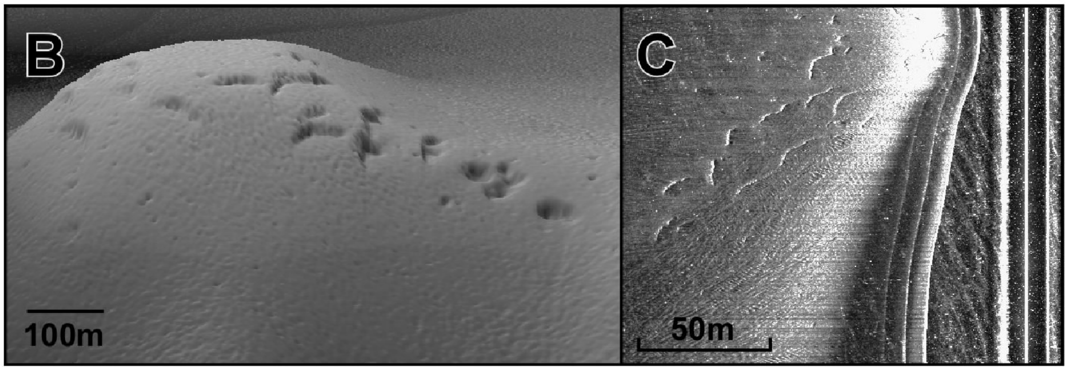
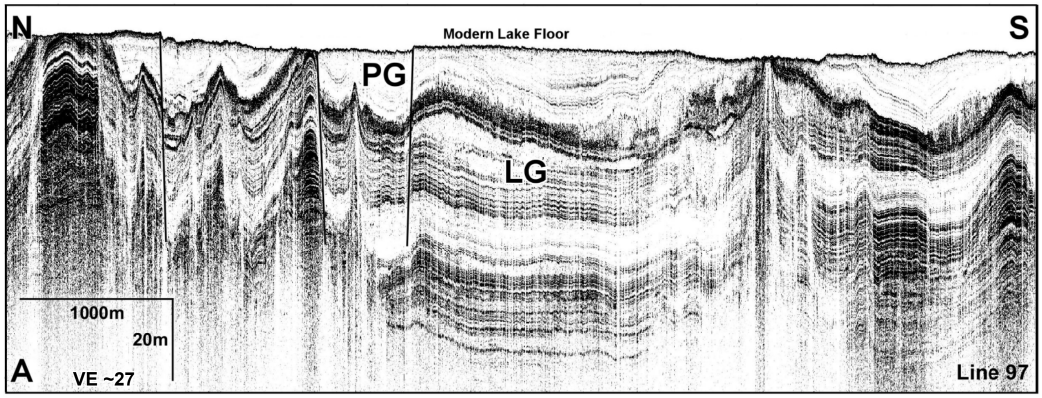


Figure 4. B. Multibeam mapping and C. side-scan profiling reveal that the East Shore fault is associated with open tension gashes on the floor of Lake Timiskaming (site marked with a star on A; see also Fig. 5B) despite high modern sedimentation rates, indicating ongoing neotectonic activity. Several northwest–southeast trending lineaments remain to be investigated in detail.



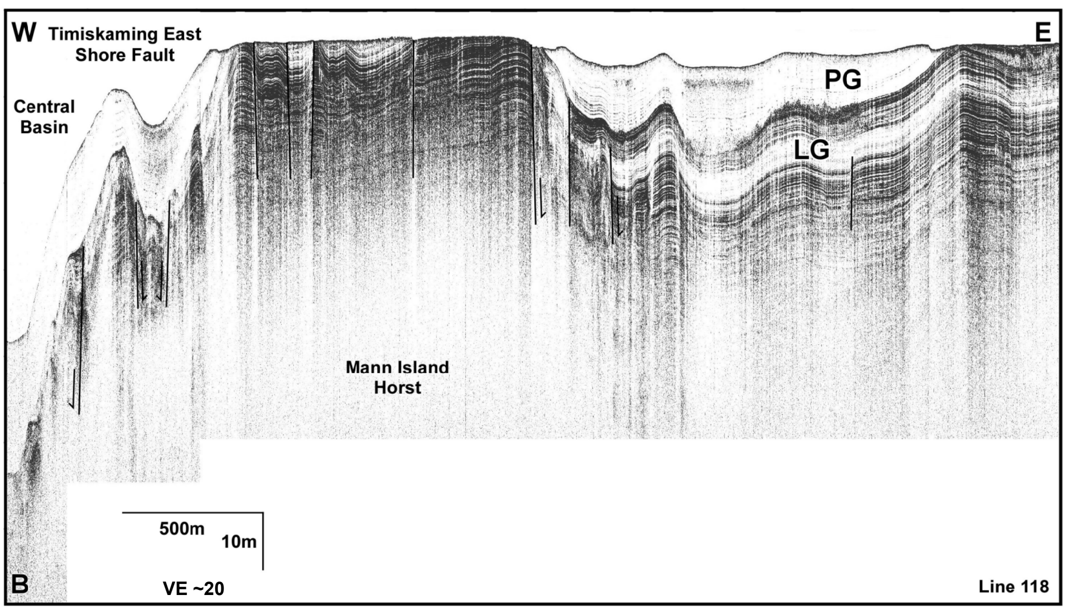
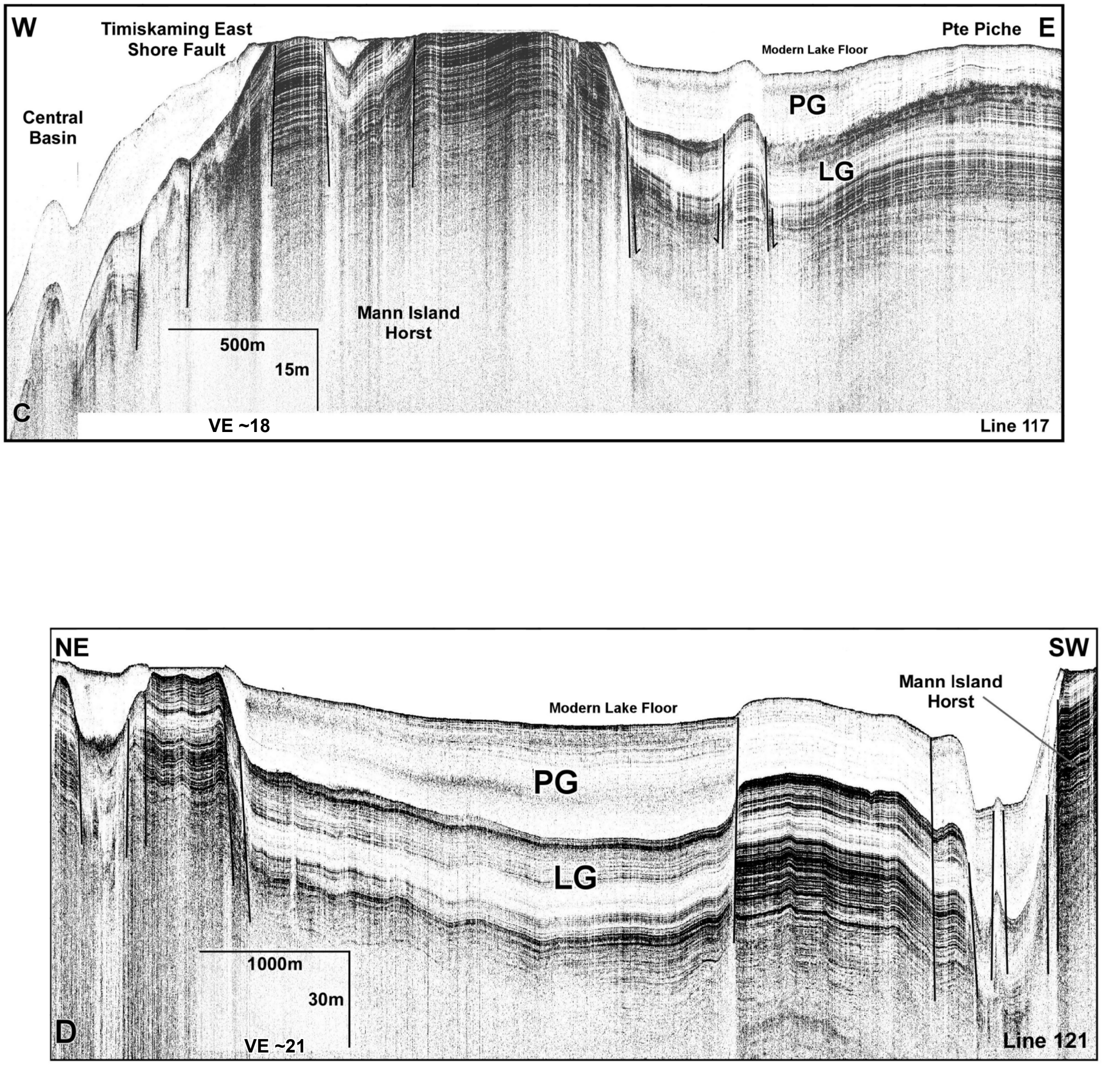


Figure 5. High-resolution seismic profiles from Lake Timiskaming (see Fig. 4A for locations) showing widespread post-depositional faulting of postglacial (PG) and late-glacial (LG) silty clays. Faults define high-standing horsts and low-standing grabens that offset the modern lake floor.



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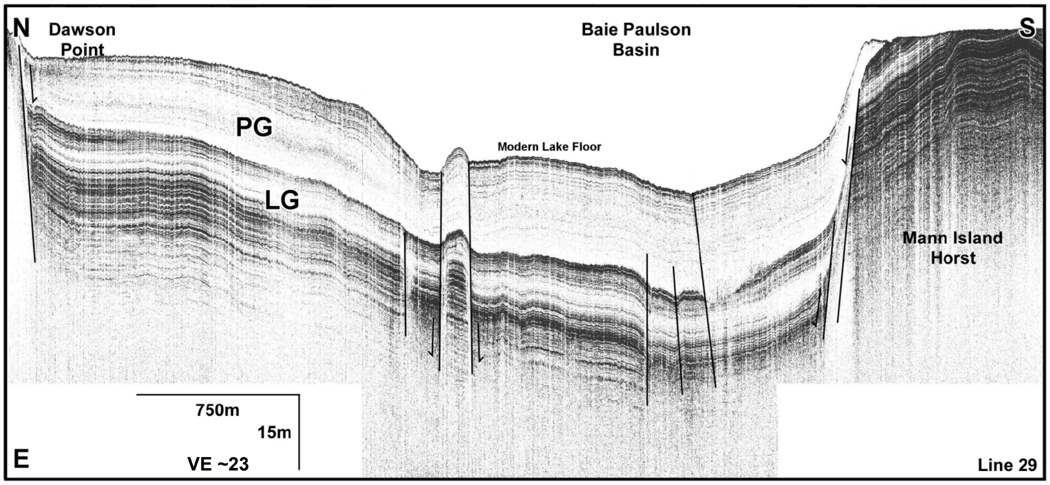

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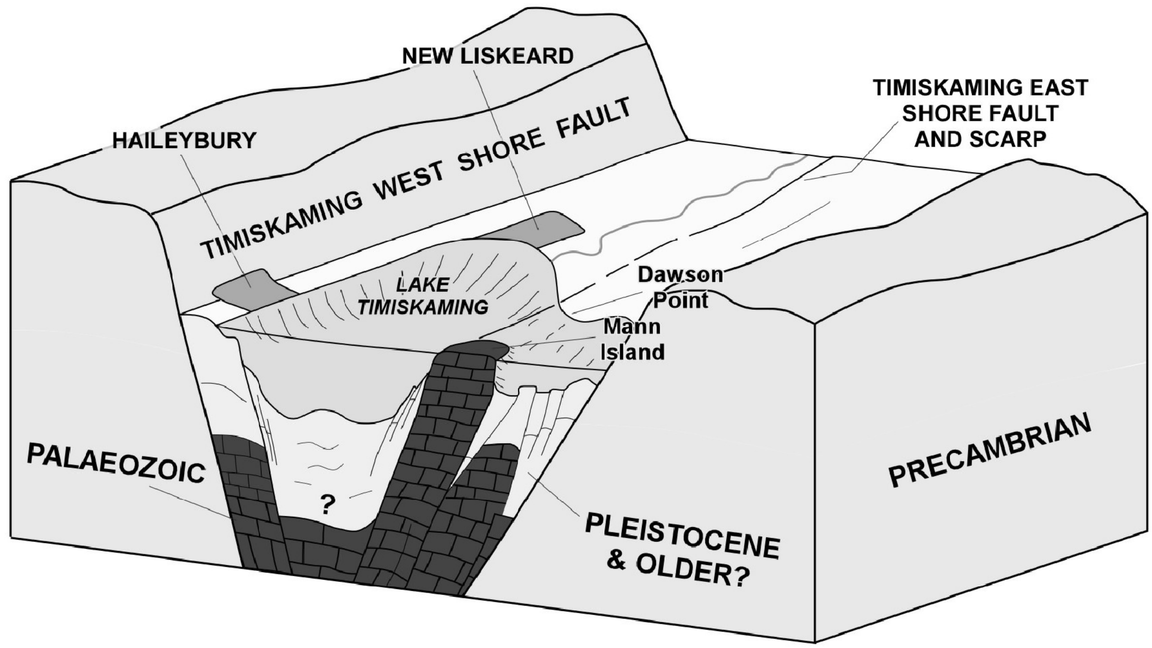

Figure 6. Simple structural model of the Timiskaming Graben as suggested by this study. Drilling is required to understand the deeper fill, which may contain a long late Cenozoic sediment and climate record.

Figure 7. Thornloe–New Liskeard fault scarp. View is looking east from point Y on Figure 4A.

Figure 8. Lighter toned 'sand blows' on the floor of the Little Clay plain northeast of New Liskeard, where liquefied sand has been injected up through overlying darker peats during seismic activity. These remain to be investigated by trenching. See Figure 4A for location.

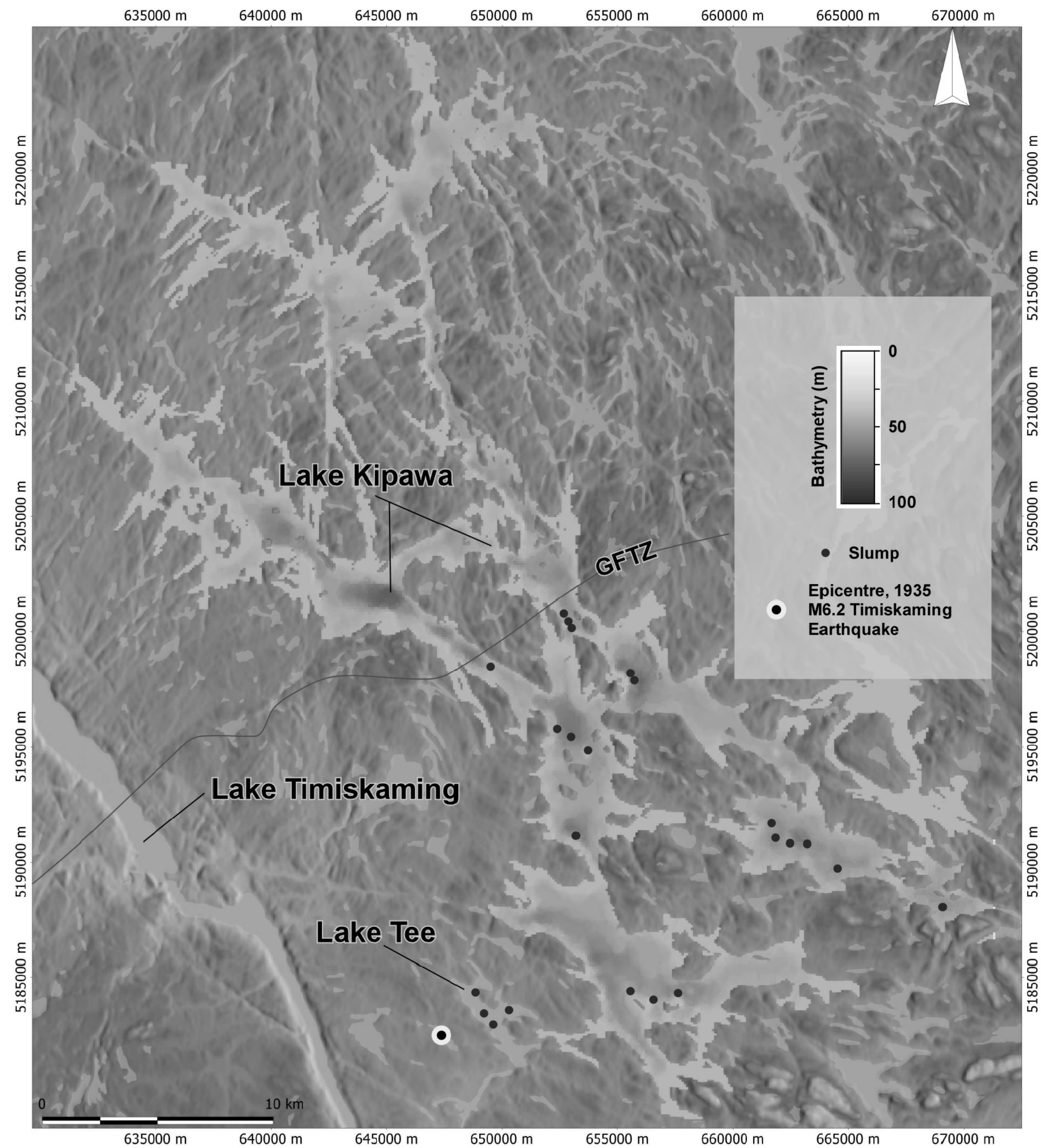

Figure 9. Distribution of landslides across the floor of Lake Kipawa produced by the 1935 Timiskaming Earthquake (M 6.2). Modified from Doughty et al. (2010). Lac Tee slumps after Shilts (1984).

Chapter 4

Recent neotectonic faulting of lake floor sediments in an intracratonic rift and the implications for regional seismicity: Timiskaming Graben, Canada

Doughty, M., Eyles, N. and Eyles, C. (2013) High-resolution seismic reflection profiling of neotectonic faults in Lake Timiskaming, Timiskaming Graben, Ontario-Quebec, Canada. Sedimentology, v 60, p. 983-1006.

ABSTRACT

The Timiskaming Graben is a 400 km long, 50 km wide northwest-trending morphotectonic depression within the Canadian Shield of eastern North America and experiences frequent intraplate earthquakes. The graben extends along the border of Ontario and Quebec, connecting southward with the Nipissing and Ottawa-Bonnechere grabens and the St. Lawrence Rift System which includes a similar structure underlying the Hudson Valley of the eastern USA. Together they form a complex failed rift system related to regional extension of North American crust during the breakup of Rodinia and later, Pangea. The Timiskaming Graben lies within a belt of heightened seismic activity (Western Quebec Seismic Zone) with frequent moderate magnitude (> M5) earthquakes including a M6.2 in 1935. These threaten aging urban infrastructure built on soft glacial sediments; postglacial landslides along the Ottawa Valley suggest earthquakes as large as M7. The inner part of the Timiskaming Graben is filled by Lake Timiskaming, a large 110 km long postglacial successor to glacial Lake Barlow that was ponded by the Laurentide Ice Sheet 9,500 years ago. The effects of frequent ground shaking on lake floor sediments was assessed by collecting more than 1000 line kilometres of high-resolution ‘chirp’ seismic profiles. Lateglacial Lake Barlow glaciolacustrine and overlying postglacial sediments are extensively deformed by extensional faults that define prominent horsts and grabens; multibeam bathymetry data suggest that faults influence the morphology of the modern lake floor despite high sedimentation rates and indicate recent neotectonic defor-

mation. The Lake Timiskaming area provides evidence of postglacial intracratonic fauling related to recurring earthquake activity along a weak spot within the North American plate.

4.0 Introduction

Eastern North America experiences moderate intraplate earthquakes that are not well understood but are a risk to several large urban centers such as New York and Montreal, several nuclear power facilities and mines, and other infrastructure sited along the eastern continental margin of Canada and the US (Kelson et al., 1996; Wu, 1998; Daneshfar and Benn, 2002; Faure et al., 2006; Stein and Mazzotti, 2007; Ma et al., 2008). Seismic risk assessment is compromised however, by lack of knowledge of earthquake recurrence intervals and magnitudes. High-resolution seismic sub-bottom surveying of lake sediments is used to expand knowledge of postglacial seismic activity beyond the limitations of relatively short instrumental and historic records (e.g., Shilts et al, 1989; Syvitski & Schafer, 1996; Oullet, 1997; Van Rensbergen et al., 1998; Wallach et al., 1998; Karlin et al., 2004; Doughty et al., 2010a, 2010b).

The purpose here is to describe the results and broader significance of a high-resolution geophysical survey of Lake Timiskaming[1], a large 110 km long waterbody that fills part of the Timiskaming Graben along the border of Ontario and Quebec in eastern Canada (Figs. 1, 2, 3). The Timiskaming district occurs within the seismically-active

[1] Note on place name spelling. The study area is variably spelled Témiscamingue, Timiskaming, Témiskaming and Témiscaming in Ontario and Quebec. Timiskaming is used throughout this paper.

Western Quebec Seismic Zone (WQSZ) which is of international concern because it extends 600 km from Lake Timiskaming southward along the Ottawa-Bonnechere Graben in Canada to Lake Champlain and the Hudson Valley in the USA and includes the New York metropolitan area (Fig. 1). In Canada, the WQSZ includes the national capital of Ottawa-Hull, and the cities of Montreal and Cornwall. Moderately large damaging earthquakes occurred in 1732 (Montreal; M5.8), in 1935 (Timiskaming; M6.2) and in 1944 (Cornwall, Ontario - Massena, New York; M5.6). The last was Canada's costliest with considerable impact on urban infrastructure. Larger multiple postglacial earthquakes up to M7 are argued to have triggered several landslides involving glaciomarine clays along the Ottawa Valley (Aylsworth et al., 2000).

4.1 Aims of this Study

Given the international importance of expanding knowledge of the history of intraplate earthquakes beyond short historical and instrumental records in eastern North America, Lake Timiskaming is of special interest as it lies within 10 km of the epicenter of the 1935 Timiskaming Earthquake (M6.2), the third largest on record in eastern Canada (Fig. 3). The lake fills a complex graben basin recently linked to ongoing crustal deformation within the North American plate (Mazzotti, 2007). The primary objective of this study was to identify the effect of recurring seismicity and regional crustal deformation on lake floor sediments of Lake Timiskaming. Doig (1991) reported deformed sediment in short (~100 cm) cores from the lake that he attributed to earthquake shaking during the 1935 earthquake. Nearby Lac Tee and Lake Kipawa (Fig. 2) lie close to the 1935 epicentre and their floors contain numerous landslides thought to be triggered by the same earthquake (Shilts, 1984; Doughty et al., 2010a). Shilts (1984) had earlier identified faults that affected both lateglacial and modern sediments in the northern part of Lake Timiskaming; he attributed these to postglacial (neotectonic) deformation but also suggested that faulting might be the product of 'glaciotectonic' sediment collapse over downwasting blocks of dead ice trapped along the floor of the lake at the close of the last glaciation. For the purposes of this study, a detailed investigation of Lake Timiskaming was conducted using multibeam bathymetry and seismic reflection data to identify the nature of deformation structures below the floor of Lake Timiskaming and to differentiate neotectonic from glaciotectonic structures.

4.2 Geological and Structural Setting

Lake Timiskaming fills part of the northwest-trending Timiskaming Graben (TG) which is part of a large 30,000 km2 extensional structure within the Canadian Shield of northeastern Ontario and western Quebec (Fig. 1). The graben forms a conspicuous 50 km wide morphotectonic depression bounded by the Montreal River and Quinze Dam faults (Fig. 2, 3). To the south, TG joins the Nipissing Graben and the Ottawa-Bonnechere Graben before connecting to the St. Lawrence Rift (SLR: Kumarapeli and Saull, 1966) that underlies the valleys of the St. Lawrence and Hudson rivers (Fig. 1). The entire system is a complex failed rift reactivated during several episodes of Rodinian and Pangean breakup over the last 700 million years. The TG may also be a much older structure possibly related to the Paleoproterozoic 'Cobalt Embayment' a failed rift arm that extended into the southern margin of the Superior Province about 2.4 billion years (Ga) ago (Rice, 1988; Jackson & Fyon, 1992; Dix & Molgat, 1998). Feeder dikes for swarms of Nipissing Diabase intrusions (c. 2.2 Ga) are controlled by the same faults that bound the graben, indicating long-lived structures. Their continued influence is also indicated by clusters of kimberlite pipes emplaced between 134 and 155 Ma ago and thought to be related to movement of the North American plate over the Great Meteor hotspot (Heaman & Kjarsgaard, 2000; Zurevinski et al., 2008). In the southern part of the Lake Timiskaming basin, the Grenville Front Tectonic Zone (GFTZ: Easton, 1992; Figs. 2, 3) is an area of high ground that marks the sharp northern boundary of the mid to late Proterozoic Grenville Province where it abuts the southern limit of the Superior Province which is composed of Archean and early Proterozoic rocks.

Large-scale post-Paleozoic subsidence along the Timiskaming Graben is recorded by downfaulted outliers of Ordovician-Silurian limestones within the graben and which formerly covered much of the surrounding Canadian Shield (Fig. 4). These occur as several westward-dipping, down-dropped blocks in sharp fault contact with the adjacent Canadian Shield bounded along the Timiskaming West Shore Fault (TWSF) and the Blanche River Fault (Hume, 1925; Bolton and Copeland, 1972; Lovell & Frey, 1976; Russell, 1984; Johnson et al., 1992). Currently, the entire Timiskaming area is being uplifted at a rate of about 4 mm a year as a consequence of glacioisostatic recovery following the last

(Wisconsin) glaciation that ended c. 10,000 years ago (Wu and Mazzotti, 2007).
Lake Timiskaming lies along the axis of the Timiskaming Graben; it is 110 km long, with a surface area of 308 km2 and mean and maximum water depths of 122 m and 209 m respectively (Figs. 2, 5, 6). The lake is 14 km wide at its northern end funnelling southward into a narrow (~ 1 km) deep 'fiord lake' basin (e.g., Eyles et al., 1991) with steep side slopes. In its overall shape from north to south, the lake basin resembles the butt, stock, trigger guard, and barrel of an upended rifle with a bayonet affixed to its extreme southern end (Fig. 3). This shape is the product of a very close structural control on the lake's shoreline configuration by faults. For some 35 km between the community of New Liskeard at its northern extremity and Baie Verhelst in the south (Fig. 3) the lake's western shoreline is an abrupt, markedly linear, 80 m high rock cliff that marks the trace of the Timiskaming West Shore Fault (TWSF; Fig. 4). Here, Lake Timiskaming consists of a narrow (between 1 and 3 km wide), deep (~ 80 m) central basin defined by parallel, fault controlled margins (Figs. 4, 5, 6).

In the northeastern part of the lake, at Mann Island, the basin splits into two, with the Paulson Bay Basin trending northeast (Fig. 4). This part of the basin is fringed by shallow (~ 5m), flat-floored bays, such as Wabi Bay, Paulson Bay, Martineau Bay, Paradis Bay, Baie Joanne, Bay Faure, Baie des Peres, Baie L'African and Baie Lavallee (Fig. 3). Upstanding blocks of Paleozoic limestone form islands (College and Mann; Fig. 4). Just north of Mission Point (in the vicinity of College Island; Fig. 5A, 6) the lake shows several enclosed 'crater-like' basins where water depths increase to 152 m; the lake attains its maximum depth of 209 m just south of Grand Campment Bay (Fig. 5C).

At Baie Verhelst the lake switches orientation to a more southerly trend for some 20 km to the mouth of the Montreal River (Fig. 3, 5B). Further down basin the lake changes orientation once again to the southeast forming a narrow straight channel bounded by the Montreal River and Cross Lake Faults (Fig. 3). A final change in orientation occurs at Chemal Opimica where the lake resembles a narrow 'bayonet' and essentially becomes a shallow 'river lake' flowing through a narrow fault-controlled bedrock trench before discharging to the Ottawa River.

4.3 Glacial History

The Timiskaming district has probably been covered by ice sheets many times in the past 2.5 million years but the geomorphic and sediment record is limited to the final stages of the last (Wisconsin) glaciation (Dyke, 2004). The region became ice free just after 10,000 years before present (ybp) when the Laurentide Ice Sheet margin withdrew north leaving an extensive belt of hummocky moraine - the Lake McConnell Moraine (Figs. 3, 7) on top of a broad northeast-trending bedrock ridge formed by the GFTZ (Boissonneau, 1968; Veillette, 1986a, b). This ridge split the ice sheet margin into two lobes (Fig. 7). A deep proglacial lake (glacial Lake Barlow) progressively filled the Timiskaming Graben as the ice sheet pulled away northwards from the moraine (Veillette, 1994). The Lake McConnell Moraine defines the southern limit of glacial Lake Barlow glaciolacustrine sediments in the Timiskaming Graben (Vincent and Hardy, 1979; Lewis and Anderson, 1989) and along the floor of Lake Timiskaming. Rising bedrock along the Cross Lake and Quinze Dam Faults formed the western and eastern shorelines of this very large proglacial lake (Fig. 2). Today its exposed floor forms an extensive plain (Little Clay Belt; Boissonneau, 1966, 1968; Morton et al., 1979; Veillette, 1986b, 1994; Rodrigues & Vilks, 1994) underlain by laminated silty-clays. A second (younger) proglacial lake (Ojibway) formed well north of the Timiskaming area and is shown in some reconstructions as being continuous with glacial Lake Barlow (glacial Lake Barlow-Ojibway; Cronin et al., 2008) but in others as a separate waterbody (Barnett, 1992). Regardless of the precise paleogeography of the lake(s), the phase of lateglacial ponding and glaciolacustrine sedimentation in the Timiskaming Graben was very brief. The lifespan of glacial Lake Barlow has been estimated to be as little as 800-900 years (Veillette, 1994) and only 2100 years for a combined Lake Barlow-Ojibway (Richard et al., 1989; Vincent, 1989; Veillette, 1994). Final drainage of Lake Barlow-Ojibway occurred at about 8000 ybp (Hillaire-Marcel et al., 1981; Vincent, 1989). Some 114,000 km^3 of lake waters were rapidly released into Hudson Bay and the North Atlantic possibly triggering the 8200 ybp cold event recorded in Greenland ice cores (Dyke, 2004; Cronin et al., 2008).

Modern sedimentation in Lake Timiskaming is seasonally controlled in response to severely cold winters. The lake is entirely ice covered during winters followed by spring breakup and the inflow of runoff from snow melt. At this time, large influxes of muddy sediment enter the northern lake basin from Wabi Creek and the Blanche and Bryson rivers that are eroding glaciolacustrine deposits of the Little Clay Belt. Muddy sediment is moved down lake by turbidity currents (Sallenave & Barton, 2006) with a marked down lake gradient in turbidity from 26.0 N.T.U in the north at Wabi Bay to 6.6 N.T.U at the southern end of the lake during peak spring runoff in June (Zettler & Carter, 1986). An annual deep water deposition rate of about 0.5 cm/yr1 was inferred for the northern part of the lake by both Shilts (1984) and Doig (1991).

4.4 Geophysical Methods

A high-resolution ‘chirp’ seismic reflection survey of Lake Timiskaming generated more than 1000 km of track line data. An EdgeTech high-resolution X-STAR digital sub-bottom profiling system was employed with a SB-216S tow vehicle (‘fish’). The system transmits an FM sonar ‘chirp’ pulse created by linear sweeping over the frequency range of 2-12 kHz for 20 ms (see Eyles et al., 2003). Water depths displayed on the seismic stratigraphic profiles used a constant velocity of 1450 m/s (Mullins & Eyles, 1996; Cauchon-Voyer et al., 2008). Water depths determined from seismic data closely agree with that identified by the Canadian Hydrographic Survey through a detailed multibeam survey of Lake Timiskaming using a Simrad EM 3000 Multibeam echosounder (CHS, 2005). With regard to sub bottom stratigraphy, depths below the lakes floor are considered approximate due to changing density and sediment type.

4.5 Results

High-resolution sub-bottom seismic profiling using the X-STAR system cannot penetrate the entire thickness of the sedimentary fill of Lake Timiskaming and therefore does not identify the nature of the bedrock floor of the graben. The X-Star system provides excellent seismostratigraphic data to a sub-bottom depth of ~ 60 m within which two very distinct seismic stratigraphic successions can be identified through most of the lake basin (Figs. 8-11). A lowermost highly reflective succession is overlain by a much more transparent succession that underlies the modern lake floor. This bipartite stratigraphy was first recognized by Shilts (1984) and attributed to late- and postglacial phases of lacustrine sedimentation respectively; the same stratigraphic 'couplet' of lateglacial and postglacial sediments is now widely recognized in lakes across much of the Canadian Shield (e.g., Shilts and Clague, 1992; Eyles et al., 2003; Lazorek et al., 2006 and refs therein).

The lowermost lateglacial succession in Lake Timiskaming (LG; Figs. 8-11) is at least 50 m thick in the deeper areas of the basin and is characterized by closely-spaced, well defined and high-frequency parallel reflections. These characteristics are typical of seasonally-deposited rhythmically-laminated ('varved') glaciolacustrine silty-clays deposited during deglaciation of the Laurentide Ice Sheet. These can be confidently related to deposition in glacial Lake Barlow because the same facies are widely exposed to the north of Lake Timiskaming as varved silty-clays in numerous stream cuts across the Little Clay Belt to the north of Lake Timiskaming (see Chan & Kenney, 1973, Quigley, 1983). In contrast, the uppermost semi-transparent seismic succession immediately below the modern lake floor shows much weaker acoustic returns in the form of very closely-spaced parallel reflections (PG; Figs. 8, 9) typical of postglacial sediments deposited in eastern Canadian lakes (see Shilts & Clague, 1992; Lazorek et al., 2006). The postglacial succession in Lake Timiskaming is on average between 15-20 m in thickness, being thickest (40 m) in the northern part of the basin adjacent to the mouths of the large muddy rivers entering Wabi Bay and Paulson Bay. These rivers feed muddy turbulent underflows down basin (see above) during the spring and summer. The postglacial phase of sedimentation in Lake Timiskaming began c. 8,000 ybp with the final drainage of glacial Lake Barlow (Cronin et al., 2008). In this regard, the maximum thickness of postglacial sediment (40 m) observed in northern Lake Timiskaming suggests maximum postglacial deposition rates of about 0.5 cm/yr-1 along the axial basin of Lake Timiskaming, in good agreement with the estimate of Shilts (1984). This value decreases down basin and the postglacial succession thins southwards to 5 m at Chemal Opimica in the lower part of Lake Timiskaming south of the McConnell moraine. The moraine marks the southern limit of lateglacial Lake Barlow in Lake Timiskaming; postglacial sediments rest directly on bedrock south of the moraine (Fig. 12).

4.5.1 Post depositional deformation

The bipartite stratigraphy established by seismic surveying shows large scale disturbance in the form of faults (Figs. 8-11) and smaller scale deformations resulting from postglacial landsliding recorded by large debris flow lobes at the base of side slopes (Figs. 5B, 13A). Lateglacial and postglacial sediment has moved downslope creating distinct fan like lobes with mounded topography downslope of steep back scarps.

High resolution multibeam bathymetric data reveal circular crater-like depressions on the lake floor south of College Island (Figs. 5, 6). These are partly filled by drapes of postglacial sediment that thicken into the depressions which are surrounded by deformed and collapsed Lake Barlow sediments. These crater-like sub basins are interpreted to be of 'glaciotectonic' origin as they result from the melt of large ice blocks trapped below sediments on the floor of glacial Lake Barlow which resulted in the collapse of overlying and adjacent sediment. These crater-like basins are very common on lake floors in eastern Canada because dead ice, weighed down by sediment, was frequently trapped below ice marginal lakes that formed during deglaciation as ice was downwasting in topographic lows (e.g., Eyles et al., 1987; Eyles et al., 2003). The subsequent melt of buried ice left enclosed steep-sided craters ('ice block depressions' or 'kettle basins') surrounded by chaotically collapsed lateglacial sediment (see Klassen & Shilts, 1982; Larocque, 1985; Kaszycki, 1987). Such basins are commonly blanketed and infilled by undisturbed postglacial sediment (e.g., Fig. 6). It is likely that other such ice block depressions may be present north of College Island but have been completely filled in and masked by postglacial sediment (Fig. 6). Their bathymetric expression becomes clearer southwards (Figs. 5, 6) in the direction of decreased postglacial sedimentation rates.

Elsewhere, the sediment fill of Lake Timiskaming is extensively disrupted by high-angle, closely-spaced normal faults that define multiple horsts and grabens (Figs. 8-11). Faults abruptly juxtapose postglacial sediment against lateglacial Lake Barlow glaciolacustrine sediments such as on the flanks of the Mann Island horst (Figs. 8B, 9A, 11A, B). Seismic data reveal the presence of two major faults defining the central basin of Lake Timiskaming; the Timiskaming West Shore Fault is already known (mapped by Russell, 1984) and its newly-identified counterpart to the east is named the Timiskaming East Shore Fault (Figs. 3, 4). Together they bound the narrow (< 3 km) central deep water basin of Lake Timiskaming between New Liskeard in the north and Baie L'Africain in the south. Rather than single, well-defined master faults, these fault systems consist of zones of closely-spaced normal faults that step down into the central basin (e.g., Fig. 9A). Figure 10B shows a prominent scarp on the lake floor directly above the Timiskaming East Shore Fault suggesting recent activity. This structure continues northward in the lake basin and emerges onshore where it can be traced across the exposed floor of glacial Lake Barlow (the Little Clay Belt) as a remarkably straight 20km-long, 8m-high linear scarp between Wabi Bay and Thornloe (Fig. 4). This feature is aligned directly above an abrupt northwest-trending step in the underlying bedrock surface thought to be a buried bedrock fault (Lovell & Caine, 1970; Russell, 1984; Doughty et al., 2010b).

Across large parts of the shallow water (< 10m), wave-influenced eastern embayments of Lake Timiskaming, lateglacial Barlow sediment is exposed directly at surface across the lake's floor with no continuous cover of postglacial sediment (e.g., Figs.

8B, 9A, B, 11B). In these areas, multiple narrow graben basins have selectively preserved small synclinal pockets of postglacial sediment suggesting the previous existence of a more extensive cover of postglacial sediment, subsequently disrupted by faulting and partially removed by erosion in shallow water (Fig. 8B).

Sub-bottom seismic profiling suggests that Paulson Bay is structurally controlled and forms a complex secondary graben structure that trends northeast, oblique to the main axis of the Timiskaming Graben. This structure is responsible for the distinct Y-shape of the northern part of the lake basin (Fig. 4). The Paulson Bay structure is bounded by faults along its southeast margin near Point Piche and on its opposing side along the southern shore of Dawson Point (Fig. 4). Paulson Bay can provisionally be identified as a broad (2 km) structurally-controlled depression formed by subsidence between bounding faults creating a broad 'sag basin' (Figs. 8, 9B, 10; see below). The thickest (~ 40 m) succession of postglacial sediment anywhere in Lake Timiskaming is accommodated in this basin. High resolution multibeam bathymetry data show that the relief across the modern lake floor at the mouth of Paulson Bay is controlled by the location of faults that offset deeply buried lateglacial sediment below (Figs. 8A, 14; see below).

Faults near Point Piche are oriented parallel to the coastline suggesting a structural control, and to the north may join the Blanche River Fault (mapped by Russell, 1984) that controls the location of the Blanche River (Fig. 4). The same structures continue southward through Chef Island to the northern coast of Mann Island and define the margins of a prominent horst of lateglacial sediment (Fig. 9A). The Baie Joanne basin appears to be a broad sub-graben between Mann Island and College Island (Fig. 4). Limestone islands on the eastern side of the basin (Chef, Oster, Mann, College) are identified as upstanding horst blocks; shallow bays that occur between the islands are graben sub-basins filled with lateglacial sediment (e.g., Baie Joanne; Fig. 4). These structures are also oriented north-east i.e., parallel to those identified below Paulson Bay but oblique to the main regional trend of the Timiskaming Graben. These secondary structures may be relict 'transfer faults' that accommodated strike-slip movement between the Timiskaming West Shore Fault and the Blanche River Fault (Fig. 4; see Discussion below). Horst blocks (the limestone islands on the eastern side of the basin) may have functioned as 'intrabasinal highs' (Withjack et al., 2002; Acocella et al., 2005; Zelilidis et al., 2008) bounding subsiding grabens that have preserved thick glacial and postglacial sediment. A detailed three dimensional study of the bedrock and sediment faults in and surrounding the Timiskaming basin is in progress but it is clear that shoreline orientation, the course of inflowing rivers and locally, the bathymetry of Lake Timiskaming basin are structurally controlled. It is significant that the southern limit of deformation in the sediments of Lake Timiskaming occurs at Baie L'Africain precisely where the shoreline of the lake basin is no longer controlled by the Timiskaming West and East Shore faults (Fig. 3; see below).

4.5.2 Interpretation: Age of faulting

The finding of faults in Lake Timiskaming is not unexpected as Shilts (1984) reported their presence at the mouth of Paulson Bay and around Mann Island but the discovery of their large extent across the lake floor from New Liskeard in the north, to Baie Verhelst 40 kilometres to the south, is very surprising. Interpretation of the age and origin of such faults is not straightforward and three possible models are considered here; glaciotectonic deformation at the end of the last glaciation, bathymetrically-controlled deposition on steep slopes, and neotectonic deformation related to crustal deformation along the graben. Each of these is examined in turn below.

Glaciotectonic deformation

Shilts (1984) was the first to identify ‘faulting of modern sediment’ in Lake Timiskaming (our italics; see caption to his Figure 76.10) in the area of horsts and graben structures near Mann Island but he also considered an alternative ‘glaciotectonic’ explanation. Closely-spaced, narrow horst blocks of Lake Barlow sediments (which he termed ‘pillars’; op cit., p. 576) and intervening grabens composed of postglacial sediment on the Mann Island Horst block (see Fig. 8B) were briefly proposed as a product of the differential subsidence of glaciolacustrine and early postglacial sediment deposited on heavily crevassed dead ice flooded by the waters of glacial Lake Barlow and early Lake Timiskaming (Shilts, 1984, p. 576). Sediments deposited within open crevasses deep enough to reach the glacier bed were considered to survive intact in contrast to adjacent sediment when the ice basement on which they were deposited began to melt down. Their superincumbent sediment would instead collapse opening up to narrow, steep sided sub-basins that filled with early postglacial sediment. Continued meltdown of underlying remnant ice well into the postglacial could result in further accommodation of postglacial sediments in subsiding graben sub-basins in fault contact with adjacent Barlow sediment (Fig. 15 A-C). Evidence of local glaciotectonic subsidence around remnant dead ice blocks is certainly recorded in glacial Lake Barlow sediments by isolated circular crater-like ‘kettle’ basins in the central part of Lake Timiskaming; these basins are surrounded by deformed Barlow sediments and are infilled by postglacial sediment (Fig. 6; see above); indeed, this is a very common situation in many Shield lakes across Canada (e.g., Eyles et al., 2003). On the other hand, the formation of widespread horsts and grabens by the melt of buried dead ice is not considered realistic as the melt of extensive ice masses buried under the floors of glacial lakes result in wholesale sediment collapse and the triggering of large subaqueous slumps recorded by chaotic deposits of debris flow and blocky mélange facies (see Eyles et al., 1987). As melt comes to an end, an irregular relief created by slumping and collapse is blanketed and smoothed by an undisturbed drape of postglacial sediment. Evidence of widespread basin-wide collapse and deformed facies cannot be identified in the fill of Lake Timiskaming. The same stratigraphic ‘couplet’ might be produced if for example, an equally short-lived phase of heightened earthquake activity argued (by some) occurred during rapid crustal rebound during deglaciation (e.g., Lagerback, 1992; Wu, 1998; Chung, 2002; Bungum et al., 2010). The end result in either case is that postglacial sediment passively drapes and infills an underlying faulted, highly deformed strata rather than forming discrete grabens of postglacial sediment faulted against horsts of Lake Barlow sediments. Shilts abandoned a ‘glaciotectonic’ interpretation for the structures in Lake Timiskaming commenting that ‘there was a good possibility' they are linked to ongoing earthquake activity (Shilts & Clague, 1992, p.1033). This interpretation is accepted here for the much wider population of faults discovered in Lake Timiskaming (see below).

Bathymetrically-controlled deposition

In any seismic study it must be recognized that apparent offsets in reflections can be generated by bathymetrically-controlled deposition in areas of strong bottom topography (Fig. 15 D). In these cases topographic highs are often 'capped' (Fig. 15 D) by hemipelagic sediment dropped from overflows, whereas the floors of topographic lows such as channels are filled with ponded turbidites; the absence of sediment deposition on intervening steeper slopes can create the impression of faulting between the two sediment accumulations on foreshortened seismic profiles. Thomas et al. (1993, Fig. 9) show a seismic profile from eastern Lake Ontario which clearly demonstrates how apparent faulting can be generated in areas of irregular bottom topography. Compression ('foreshortening') of the horizontal scale on seismic lines (in comparison with vertical thickness scales) may create the appearance of abrupt steps and faults in otherwise undulating draped successions. While it is important to recognize this effect, in general, it cannot explain the numerous closely-spaced horsts and grabens that affect both late and postglacial sediments in the Mann Island horst (Figs. 9A, 11A, B). These structures not only offset the modern lake floor but also have a distinct rectilinear planform on the lake floor (Fig. 14) typical of intersecting faults that have controlled local bathymetry. Moreover, the side slopes of these blocks on non-exaggerated seismic profiles are relatively steep (approximately 45°) which greatly exceeds the angle of repose for fine grained, wet sediment. Correspondingly, the blocks cannot simply be erosional high standing remnants resulting from channeling or erosion by wave action of surrounding postglacial and late glacial sediment (Fig. 15D). Thus horst and graben structures (Fig. 9) are not an expression of bathymetrically controlled deposition where sediment was selectively deposited in narrow steep sided depressions between upstanding blocks. These structures are demonstrably the result of faulting as suggested by Shilts (1984) after he had rejected a glaciotectonic model.

Neotectonic deformation

Many lines of evidence support a third ('neotectonic') hypothesis that both lateglacial and postglacial sediments have been simultaneously affected by geologically recent faulting (Fig. 15 E). The topography of some areas of the modern lake floor closely mimics that found on underlying faulted sediments (e.g. Fig. 14) which by itself strongly suggests the continuing effects of through-going faults and recurring activity. Mazzotti (2007) argued that the seismically-active Timiskaming Graben is currently experiencing intraplate deformation as a distinct 'weak zone' within the North American plate. The effects of postglacial (neotectonic) faulting of bedrock have long been recognized in bedrock mines within the Timiskaming district (Miller, 1913) and there is a strong bedrock structural control on drainage patterns on the surface of the Little Clay Belt (Hume 1925). A neotectonic origin for the faults in Lake Timiskaming is furthermore, strongly indicated by the regional distribution of deformation structures in the lake basin. The southernmost extent of faulting is sharply defined and occurs at Baie L'Africain and Baie Verhelst (Fig. 3) precisely in the area where the configuration and orientation of the lake basin is no longer controlled by the strike of the Timiskaming West Shore and East Shore faults (Figs. 2, 3). This relationship indicates that these two large faults bounding the central basin of the Timiskaming Graben (Fig. (Figs. 4, 9A, 10B) have been active in the very recent past; postglacial faults are not present in the lake basin to the south where these two regional faults no longer control the configuration of the lake basin. This geographic relationship between specific bedrock faults and the spatial extent of faulting in the sediment fill of Lake Timiskaming is telling and strongly supports a neotectonic model. This relationship would not be expected in the case of the other two hypotheses considered. The presence of a long linear scarp in Barlow sediments right above the onshore continuation of the Timiskaming East Shore Fault (Fig. 10) has also been attributed to neotectonic movement of the underlying bedrock structure (Doughty et al., 2010b).

Neotectonic faulting within the Timiskaming Graben is further supported by GPS data published by Mazzotti (2007) that identify differential velocities and trajectories of that part of the Western Quebec Seismic Zone in the immediate vicinity of the Timiskaming Graben (Hayek et al., 2008; Fig. 1). This deformation is associated with ongoing seismic activity. The largest recorded earthquake in the Timiskaming area occurred in 1935 (M6.2) with its epicentre located just 10 km east of Lake Timiskaming below Lake Kipawa where the Grenville Front Tectonic Zone crosses the Timiskaming Graben (Bent, 1996; Fig. 3). Some 90% of chimneys were damaged in the area suggesting a maximum Mercalli intensity of VII (Hodgson, 1936a, b). On average, an M>3 earthquake occurs in the Timiskaming area every two years (Adams et al., 2000). Adams et al. (2000) inferred the presence of a northwest trending fault under Lake Timiskaming based on focal plane solutions for the January 1, 2001 Kipawa Earthquake; this agrees very well with the results of the seismic study reported here which point to the Timiskaming West Shore and East Shore faults as active.

Earthquake-triggered landslides very similar to those found in Lake Timiskaming (Fig. 13) occur across the floor of Lake Kipawa 10 km north of the community of Témiscaming and are clustered around the epicentre of the 1935 earthquake (Doughty et al., 2010b). Other large postglacial landslides have been mapped down valley in the Ottawa area and attributed to large (M7) earthquakes (Aylsworth et al., 2000; Moernaut et al., 2007). It is noted that Antevs (1925) mapped changes in varve thickness within glacial Lake Barlow sediments in exposures across the Little Clay Belt to identify the rate of ice melt and retreat. Significantly, he identified intraformational folding within varves exposed in long railway excavations extending over several tens of kilometres, just east of Lake Timiskaming; deformation was attributed to either landsliding or iceberg scouring (Antevs, 1925, p. 81) but this large, extensively deformed zone may also be a ‘seismite.’

In summary, a wide range of data sets including high-resolution seismic profiles (Figs. 8-11), multibeam bathymetry data (Fig. 5, 14), the configuration of shorelines, the location of rivers and the regional distribution of faults (Fig. 4), together with the presence of scarps on the lake floor (Fig. 10) and surrounding Little Clay Belt (Fig. 4) all point to neotectonic crustal deformation along the Timiskaming Graben. The Timiskaming West Shore and East Shore Faults appear to be the dominant active faults and are primary targets for further detailed geophysical investigations elsewhere along their length.

4.6 Discussion

In regard to the original objectives of this investigation, results confirm that known ongoing crustal deformation in the Western Quebec Seismic Zone (Mazzotti, 2007) is expressed as neotectonic faults that affect lateglacial and postglacial sediments on the floor of Lake Timiskaming. Some 1000 km of high-resolution 'chirp' seismic sub-bottom profiles across the lake floor reveal an extensive suite of normal faults that affect glacial Lake Barlow varved silty-clays and overlying postglacial sediments. These are not the product of the melt of buried remnant dead ice blocks but indicate recent neotectonic faulting consistent with ongoing seismicity in the Western Quebec Seismic Zone.

Results of this study greatly expand understanding of intracratonic deformation in eastern North America (see Johnson and Schweig, 1996; Kelson et al., 1996; Tuttle, 2001; Wheeler and Crone, 2001; Grant, 2002; Csontos et al., 2008 for reviews). The scale of deformation identified here is unusual and hitherto unreported and is more typical of the active western margin of North America (e.g., Karlin et al., 2004; Sherrod et al., 2004). The Timiskaming structures are more closely comparable to those in other intracratonic rifts such as the Rhine Graben in Germany, the Kleszczow Graben in Poland and others (e.g., Brodzikoski et al., 1987; Scholz et al., 1998; Gawthorpe and Leeder, 2000; Morley, 2002; Withjack et al., 2002; Derer et al., 2003; Gvirtzman 2006). It is interesting to note that the Timiskaming Graben has broadly the same dimensions as the 200 km long, 10-15 km wide Dead Sea transform rift basin (Shamir, 2006; Ben-Avraham and Lzar, 2006; Haberland et al., 2007) and seismic profiles through the fill of that basin are in fact closely comparable to those from Lake Timiskaming (see Gvirtzman, 2006; Shamir, 2006). The presence of possible transfer faults oblique to the axis of the Timiskaming Graben (see above) indicates that some strike-slip movement and horst block rotation has occurred in the past, similar to that of the Dead Sea (e.g., Storti et al., 2003; Schreurs, 2003; Shamir, 2006), during one or more Paleozoic Appalachian orogenies marking the amalgamation of Pangea (e.g., Dix and Molgat, 1998; Dix and Robinson, 2003; Dix et al., 2007). Faure et al. (1996) showed that strike-slip movement along the St.Lawrence Rift of southern Quebec occurred after the early Cretaceous rifting and separation of North America from Greenland.

An expanded program of seismic profiling of the many other large lakes in the Timiskaming district is now needed to identify the temporal record of postglacial seismic activity in the WQSZ. The need to continue this work is pressing because the St.Lawrence Rift system is thought to extend westward into the heavily-populated Lake Ontario basin with its nuclear generating facilities near Toronto (e.g., Basham and Adams, 1989; Adams and Basham, 1989). Seismic risk to large cities such as Ottawa and Montreal is compounded by the presence of glaciomarine Leda Clays deposited in the lateglacial Champlain Sea, and aging urban infrastructure (see Motazedian and Hunter, 2008; Rosett and Chouinard, 2008). Major earthquakes occurred in the area in 1732 (Montreal; M5.8) and in 1944 (Cornwall; M5.6) and as related above, earlier large postglacial earthquakes are recorded by postglacial landslides along the Ottawa Valley (Aylsworth et al., 2000). Earthquake-triggered debris flows occur in several Appalachian lakes (Shilts, 1984; Shilts et al., 1989, 1992; Turgeon et al., 1994). Locat et al. (2003) and Cauchon-Voyer et al. (2008) identified the effects of repeated earthquake activity in Saguenay Fjord within a graben-like structure offset from the St.Lawrence Rift in a structural setting very similar to the Timiskaming Graben (Tremblay et al., 2001; Fig. 1).

4.7 Conclusions

The Timiskaming Graben is a large and long-lived morphotectonic basin resulting from subsidence within the Canadian Shield of eastern Canada and the USA. It forms part of a regional failed rift system (the St.Lawrence Rift) that includes the Hudson Valley of northern USA and is one of the most seismically active areas in eastern North America (Western Quebec Seismic Zone). Several large metropolitan areas lie within its boundaries (New York, Montreal, and Ottawa). Lake Timiskaming fills part of the graben and is a large deep 'successor' lake formed by the drainage some 8,000 years ago of a regionally extensive ice-dammed lake (glacial Lake Barlow) formed during the last retreat of the Laurentide Ice Sheet. Normal faults affecting Lake Barlow varved silty-clays and overlying postglacial sediments, seen on sub-bottom profiles collected with a high-resolution 'chirp' seismic system, indicate recent neotectonic faulting consistent with ongoing seismicity in the Western Quebec Seismic Zone (and not resulting from the melt of buried, remnant ice blocks). A recent geophysical model that identifies the Timiskaming Graben as a deforming zone within the North American plate is supported by geological and geophysical investigations of sediments in Lake Timiskaming. Detailed geophysical and sedimentological studies of the type reported here have major societal relevance in areas of eastern North America affected by intraplate earthquakes.

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List of Figures

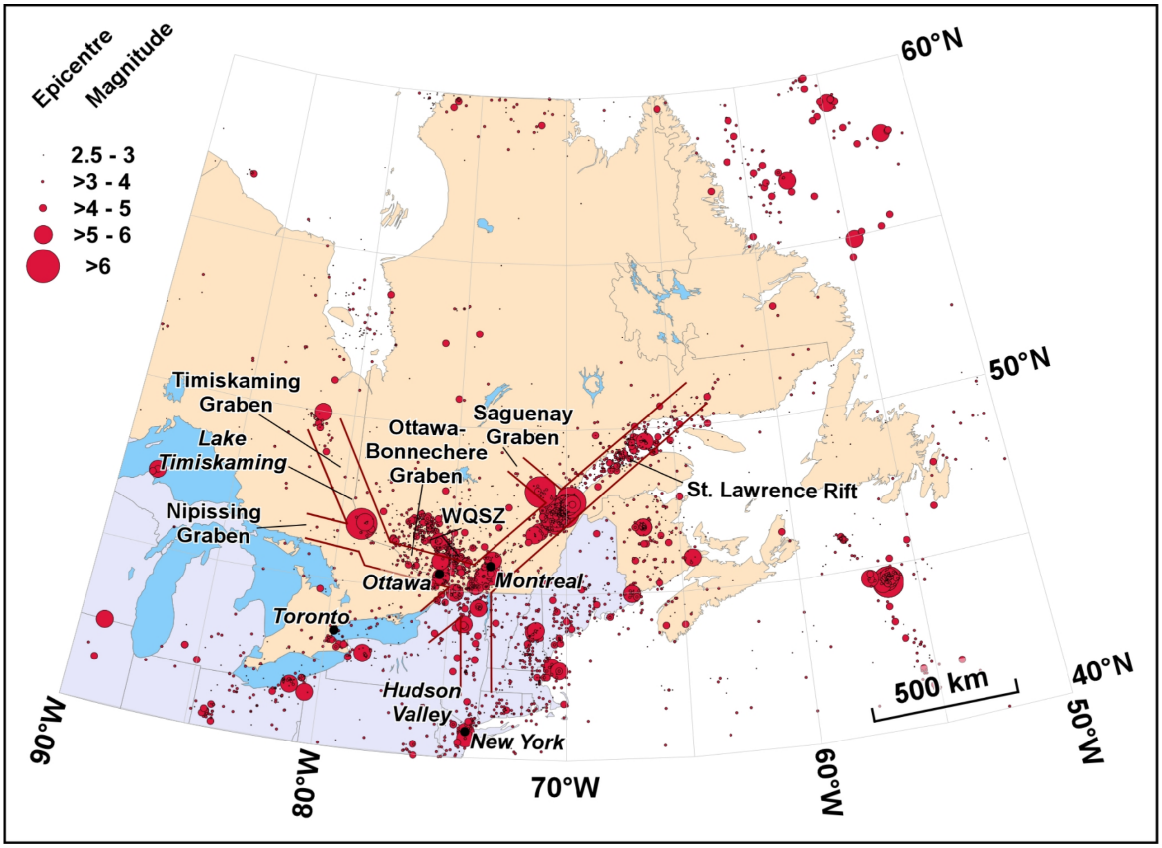

Fig. 1 Regional structural setting of the Timiskaming Graben and the St.Lawrence Rift system in eastern North America showing earthquake epicentres. Epicentres are plotted for the years 1670 through 2011; data are from Halchuk (2009) and Natural Resources Canada (2011).

Fig. 2 (Next Page) Digital elevation model of Timiskaming Graben showing (in green) broad belt of glaciolacustrine deposits of the Little Clay Belt left by drainage of glacial Lake Barlow (Fig. 7). Extent of clay belt is sharply delineated west and east by sharply rising bedrock along bounding faults of the graben. The Ontario-Quebec border is indicated by the dashed line (through the center of Lake Timiskaming). Major faults (including the Timiskaming West Shore Fault) are indicated. The Timiskaming East Shore Fault (this study) is indicated with a dashed line.

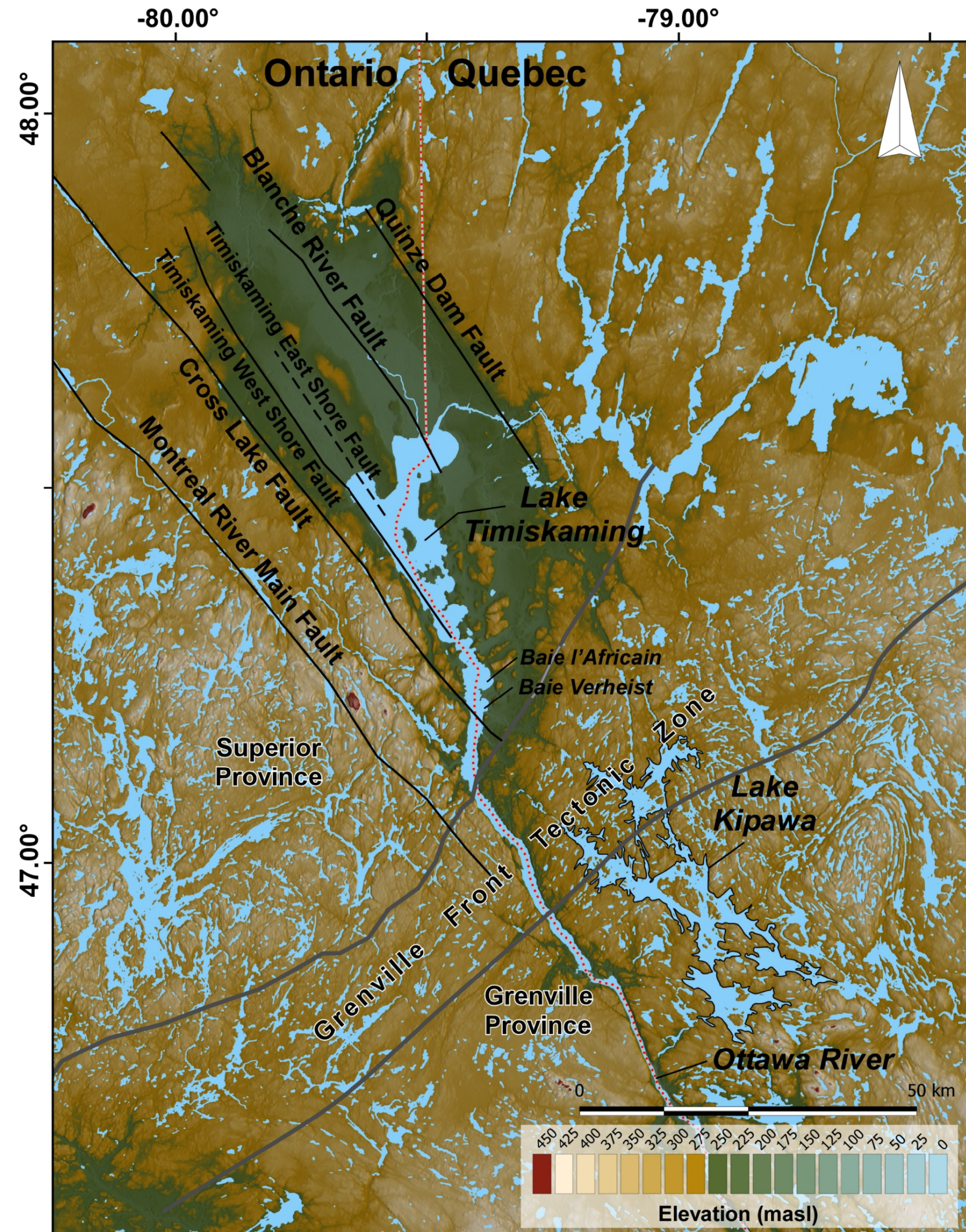

Fig. 2 Continued ...

Fig. 3 (Next Page) Simplified regional geology and principal faults associated with the Timiskaming Graben (see also Fig. 4). Neotectonic deformation in Lake Timiskaming bottom sediments is limited to the northern part of the basin above Baie Verhelst in the vicinity of the Timiskaming West and East Shore faults, and inferred faults below Paulson Bay (see Fig. 4). The Montreal River and Cross Lake faults in the southern part of the lake appear to be inactive. Insets show location of bathymetry maps (Fig. 5A-D). Coordinates shown using UTM Z17.

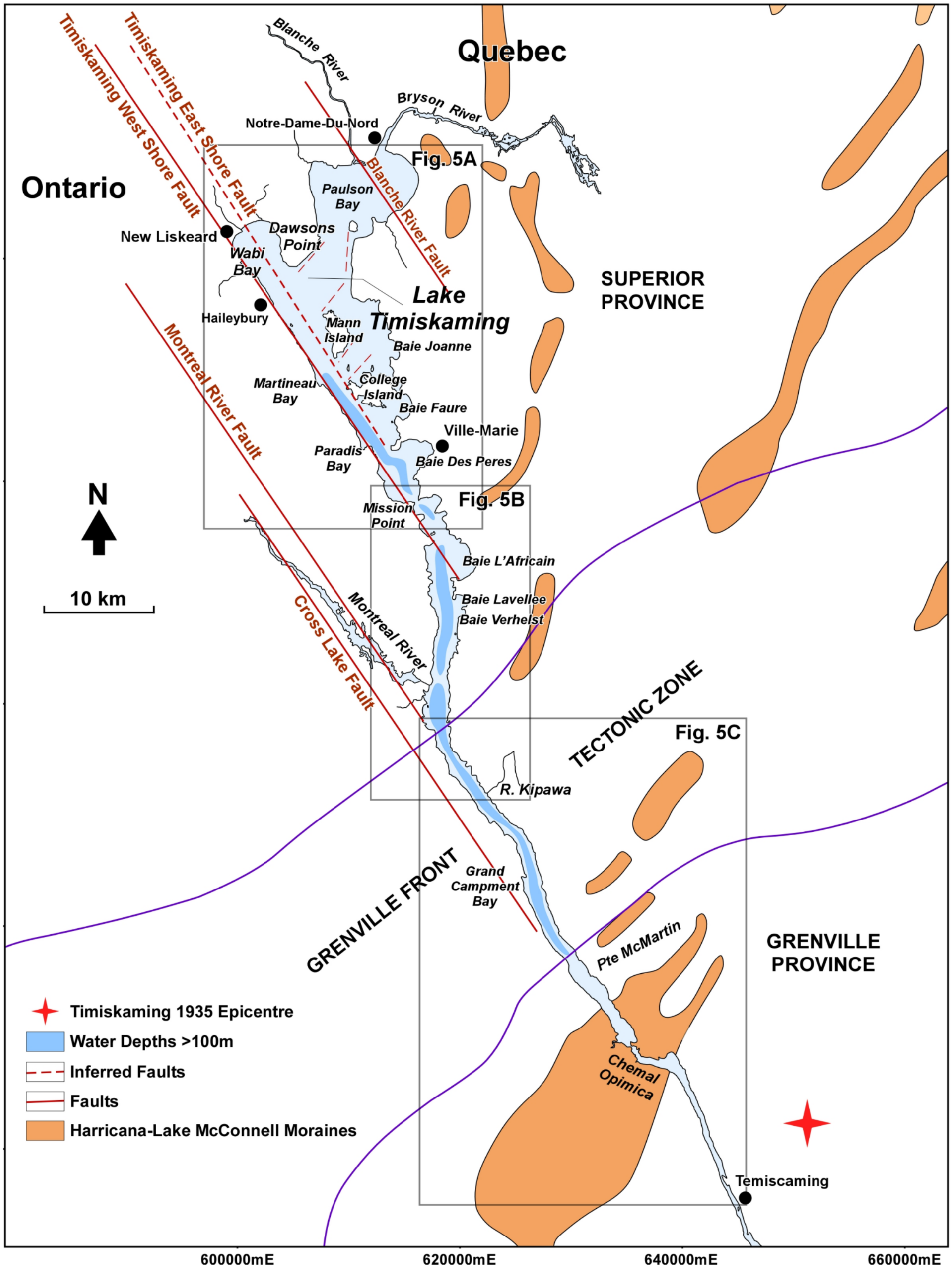

Fig. 3 Continued ...

Fig. 4 (Next Page) Simplified structural map of north Lake Timiskaming (after Russell, 1984 with additions) showing downfaulted blocks of Paleozoic limestone within the graben. Inferred structures indicated by dashed lines. The Timiskaming West Shore Fault is expressed as a prominent bedrock cliff along the lake's shore and with the East Shore Fault defines the central basin of the Timiskaming Graben. Seismic data suggest Mann Island and College Island are horst blocks; intervening bays are grabens defined by faults oriented oblique to the central graben basin. Coordinates shown using UTM Z17.

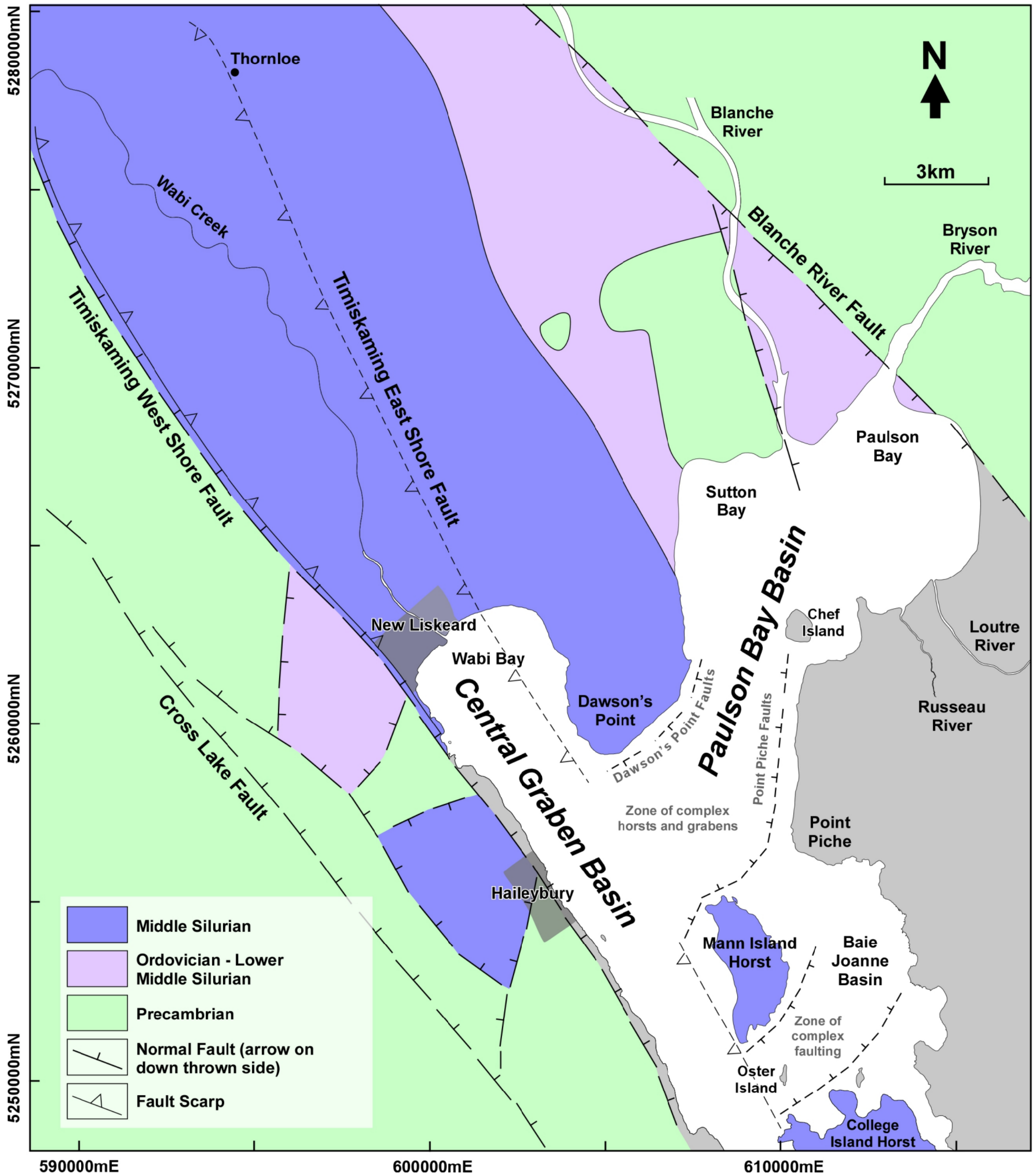

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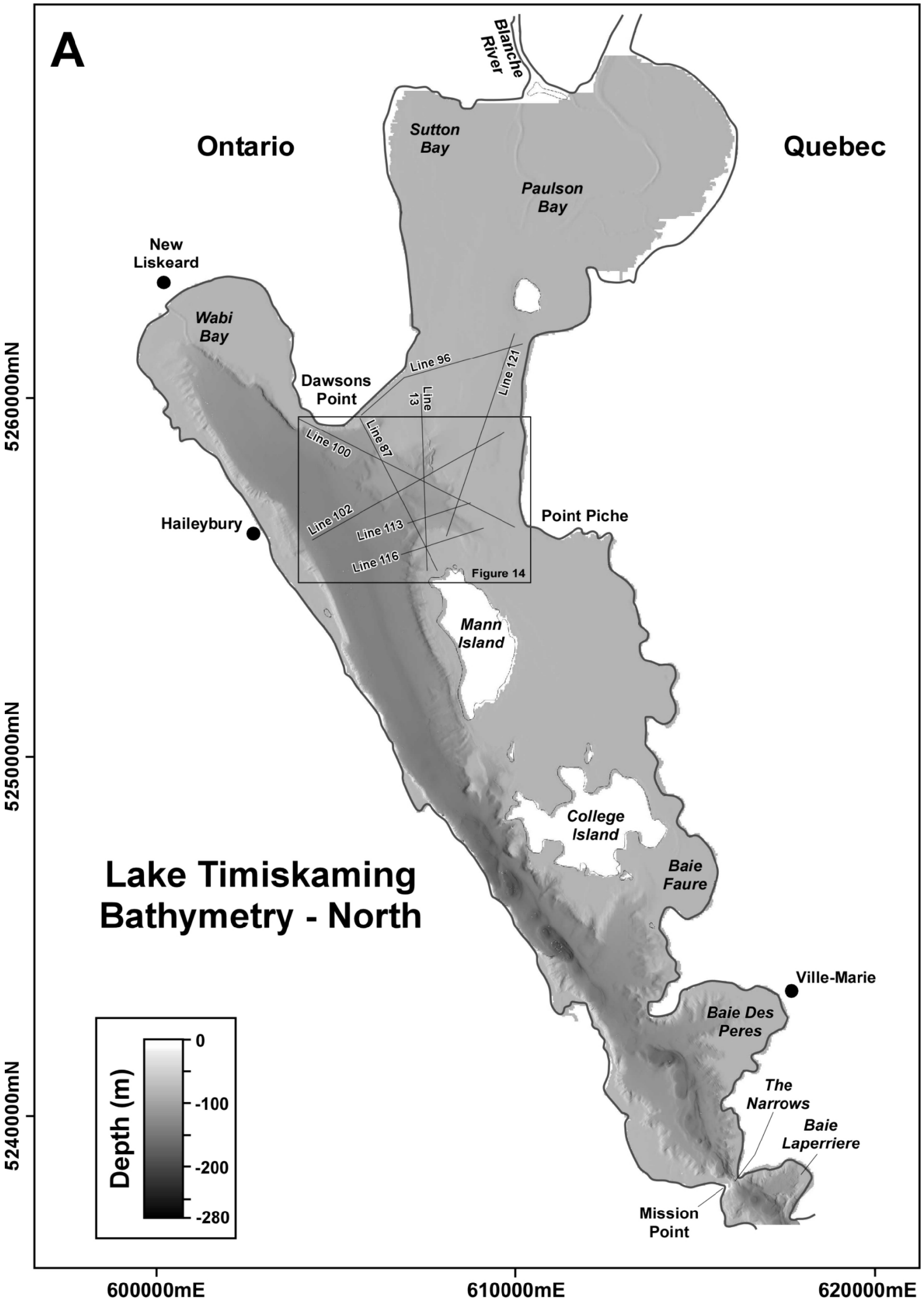
Fig. 5 (A-D; Next Page) Bathymetry of Lake Timiskaming (after Canadian Hydrographic Service, 2005) showing location of seismic survey track lines depicted on Figures 8-11. Enclosed depressions on lake's floor are lateglacial 'kettle basins' (Fig. 6B). Coordinates shown using UTM Z17. Cell resolutions of original bathymetric data are between 1m and 25m (the latter in shallow areas); data shown were re-sampled to 5m. (E) Location of all seismic lines throughout Lake Timiskaming.



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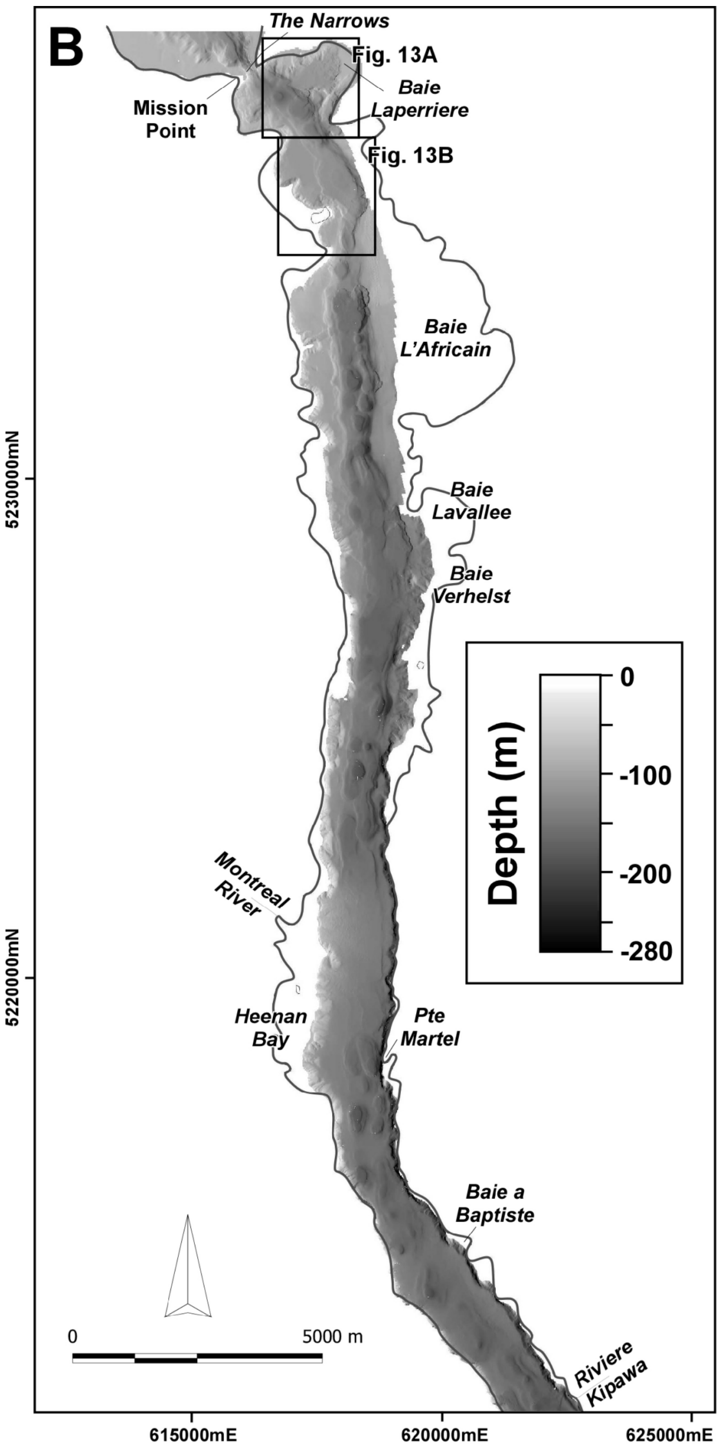

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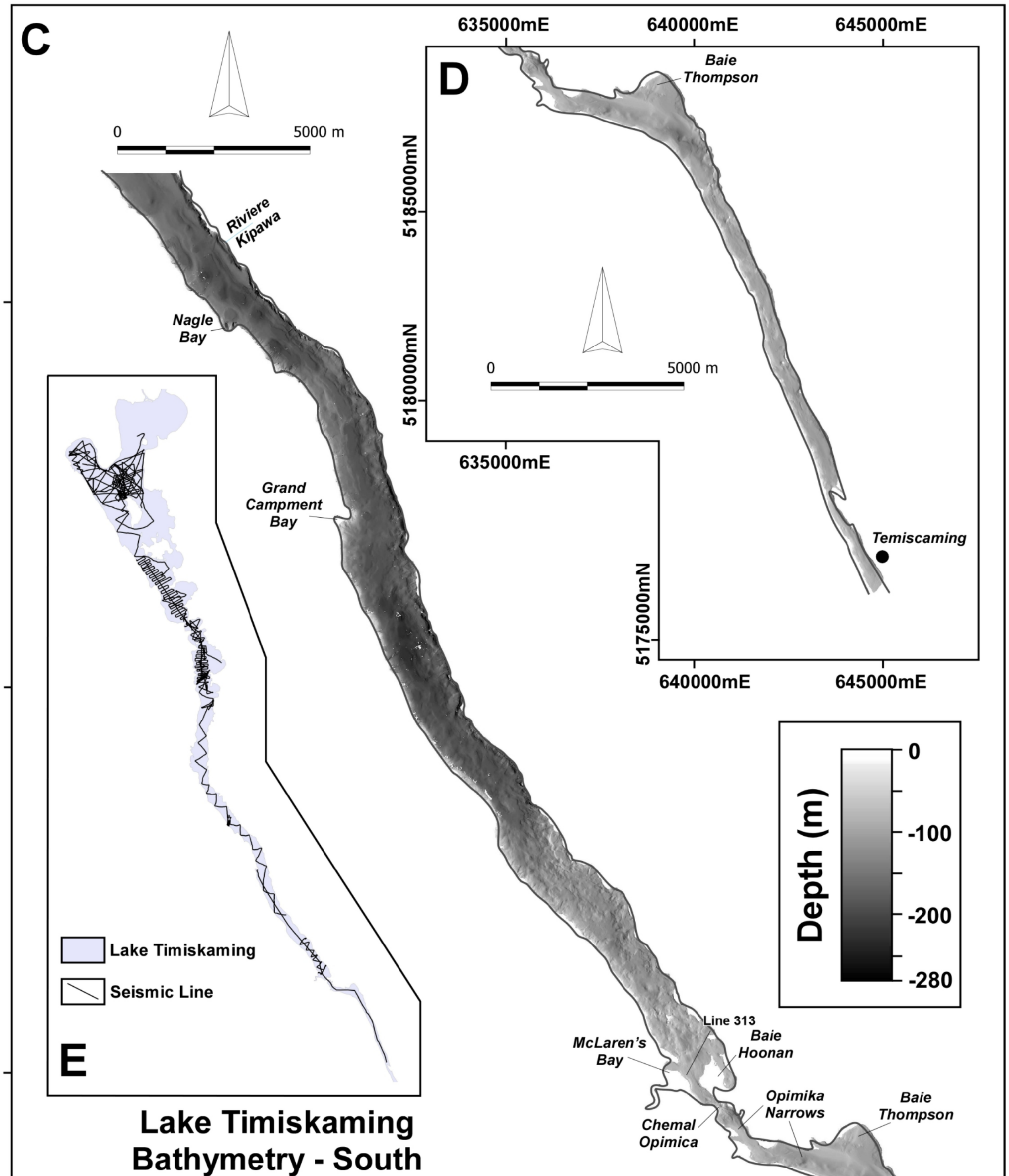

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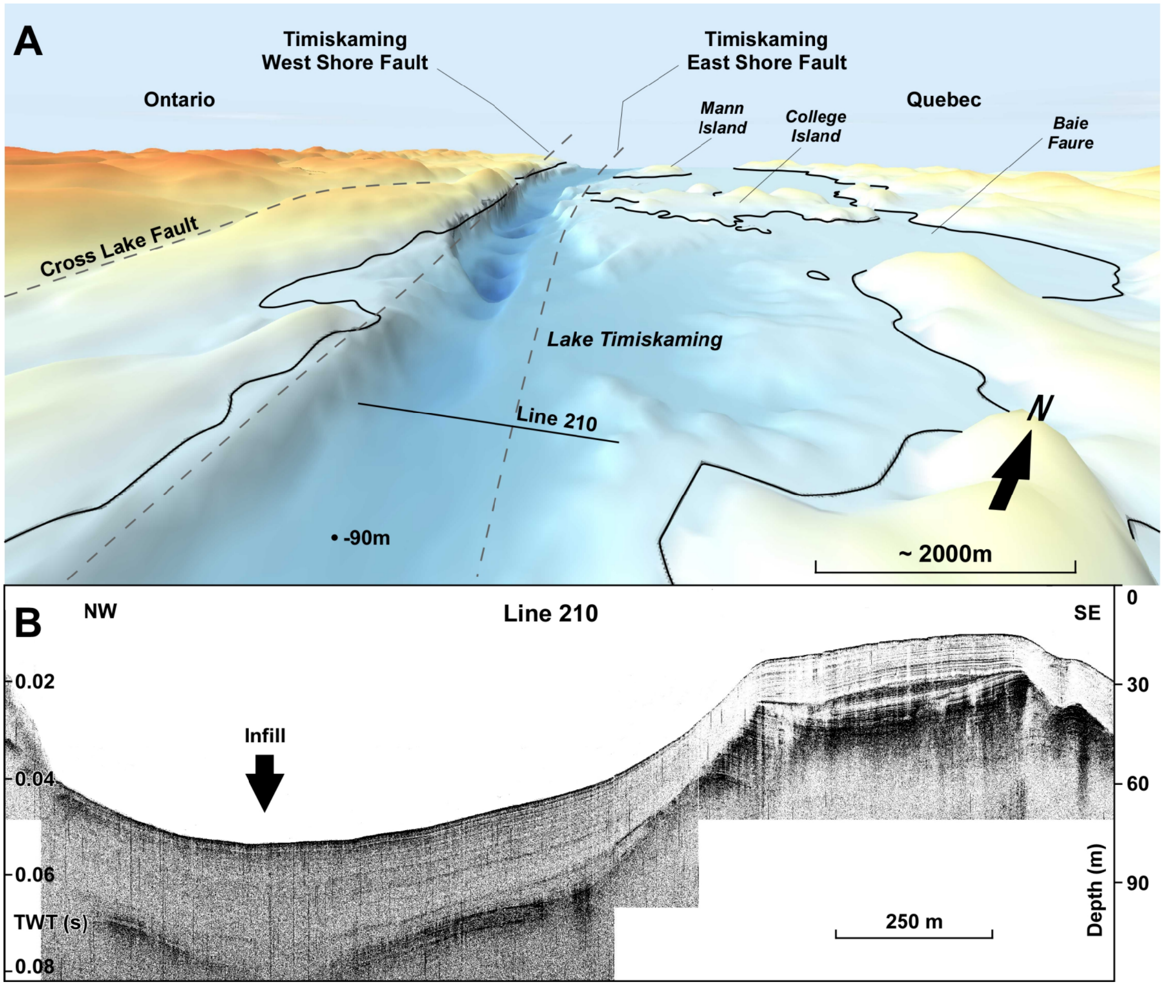
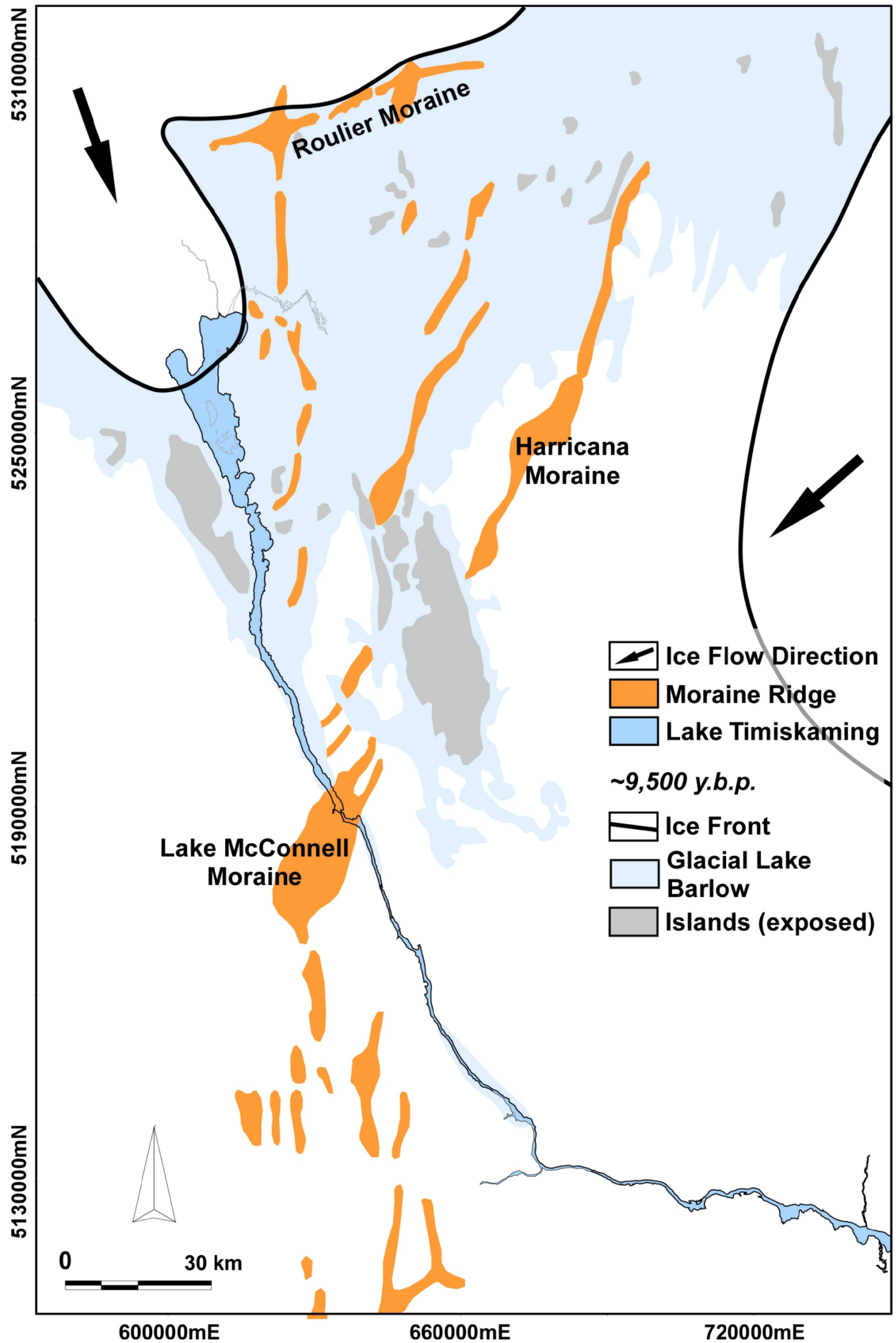


Fig. 6 (A) Northward looking aerial perspective of bathymetry of Lake Timiskaming toward New Liskeard (Fig. 4). Note steep faulted margin of lake to west, deep central basin and extensive shallow water shelf areas to east. Circular crater-like depressions on lake floor are ‘kettle basins’ resulting from the melt of buried ice at the end of the last glaciation. (B) Seismic section across infilled ‘kettle basin’ near College Island.

Fig. 7 (Next Page) Lateglacial extent of glacial Lake Barlow (light blue) at c. 9,500 ybp (after Veillette, 1986, 1994) in relation to modern Lake Timiskaming (dark blue). Lake Barlow flooded the Timiskaming Graben as the Laurentide Ice Sheet withdrew from the Lake McConnell Moraine. Its drainage left a flat clay plain (Little Clay Belt; Fig. 2) and a thick sediment fill along the floor of Lake Timiskaming. Coordinates shown using UTM Z17.



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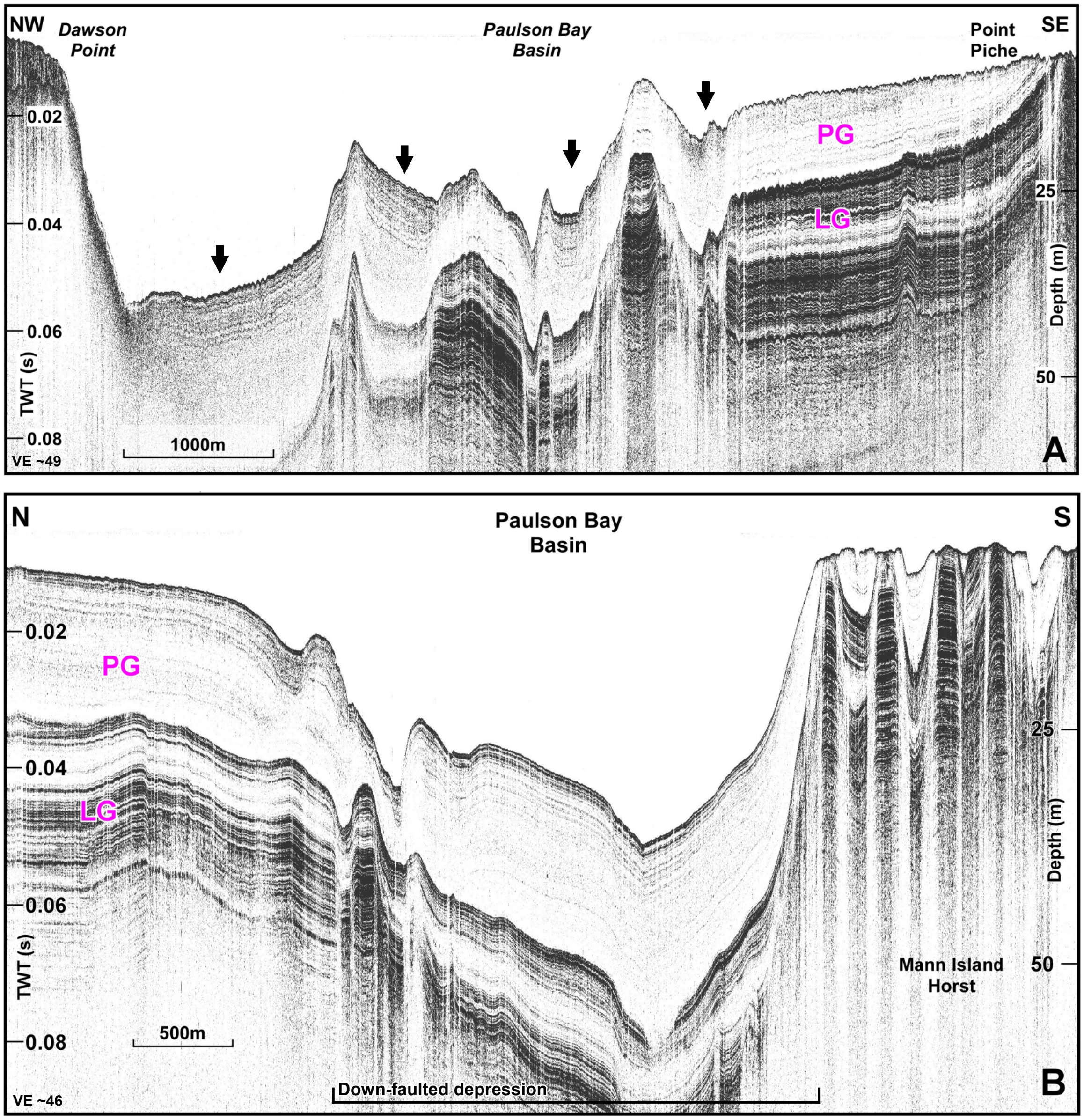

Fig. 8 (A) 7.1 km long profile (line 100: Fig. 5A) across Paulson Bay showing grabens (arrowed) and horst blocks in lateglacial (LG) and postglacial (PG) sediments. (B) 5.3 km long north-south seismic profile from Mann Island to Dawson Point across down-faulted depression (as indicated) at mouth of Paulson Bay (line 13: Fig. 5A).

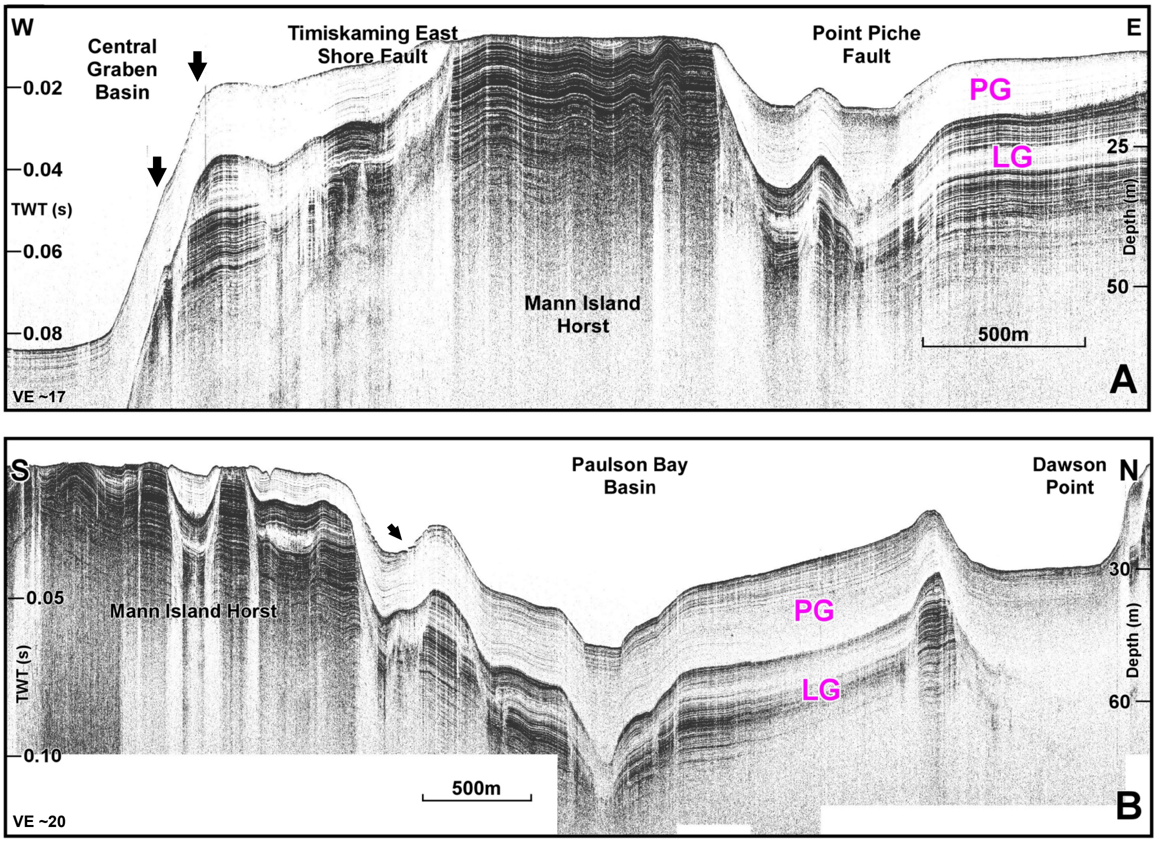

Fig. 9 (A) 3.5 km long seismic profile (line 116: Fig. 5A) across the central basin of Lake Timiskaming and the Timiskaming East Shore Fault, Mann Island horst block and Point Piche Fault zone (Fig. 5A). Area of closely-spaced normal faults indicated (approximately) by arrows. (B) 5.3 km long seismic profile (line 87: Fig. 5A) between Mann Island and Dawson Point showing fault control on bathymetry and area of open fissures on lake floor above faults (arrowed).

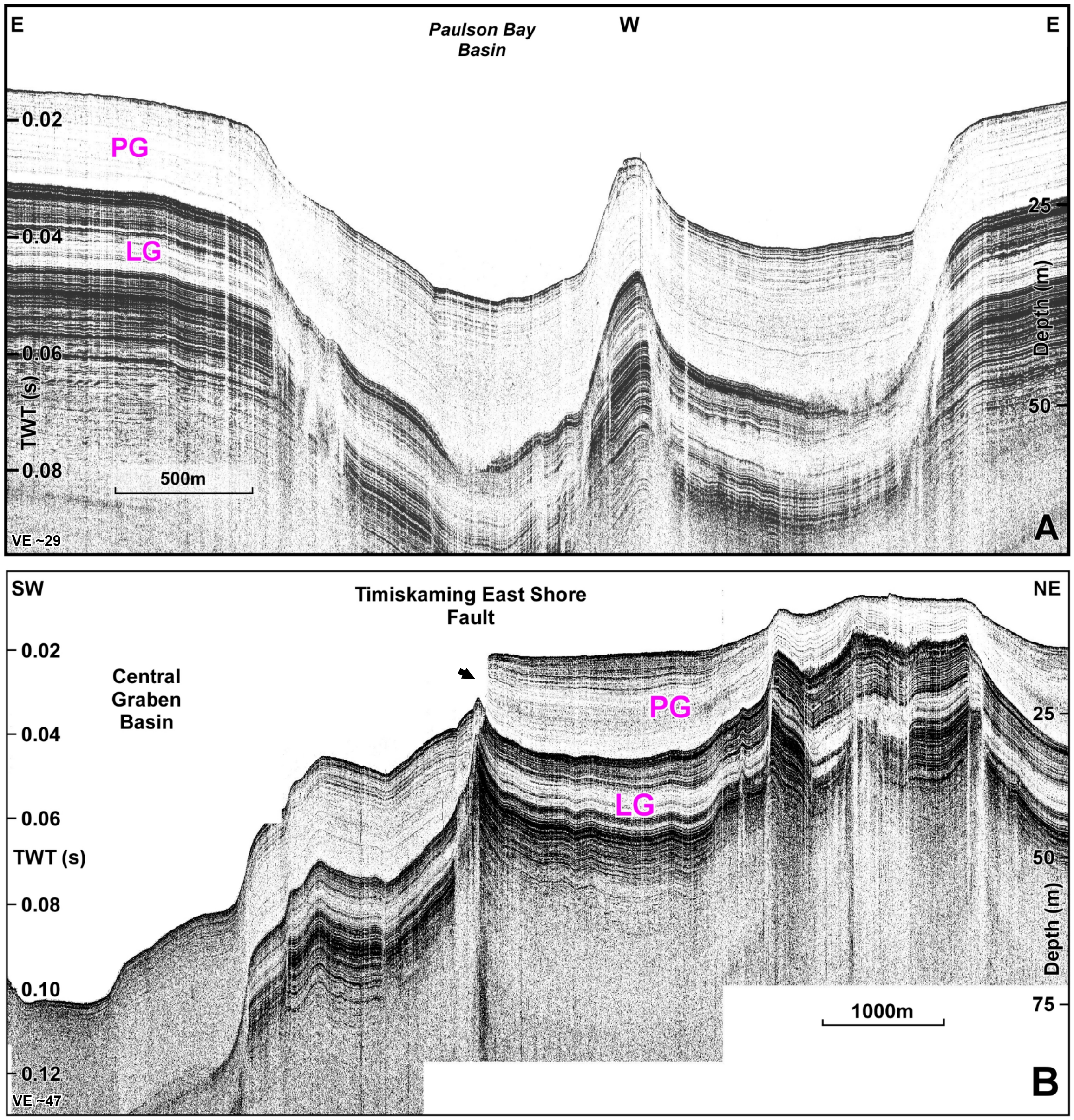

Fig. 10 (A) 3.8 km long seismic profile from Mann Island to Dawson Point across Paulson Bay (line 113: Fig. 5A). (B) 8.7 km long seismic profile showing fault scarp (arrowed) above Timiskaming East Shore Fault on floor of Lake Timiskaming in Paulson Bay Basin (line 102: Fig. 5A); height of scarp is comparable to the scarp exposed onshore to the north between Wabi Bay and Thornloe along same fault (Figs. 4).

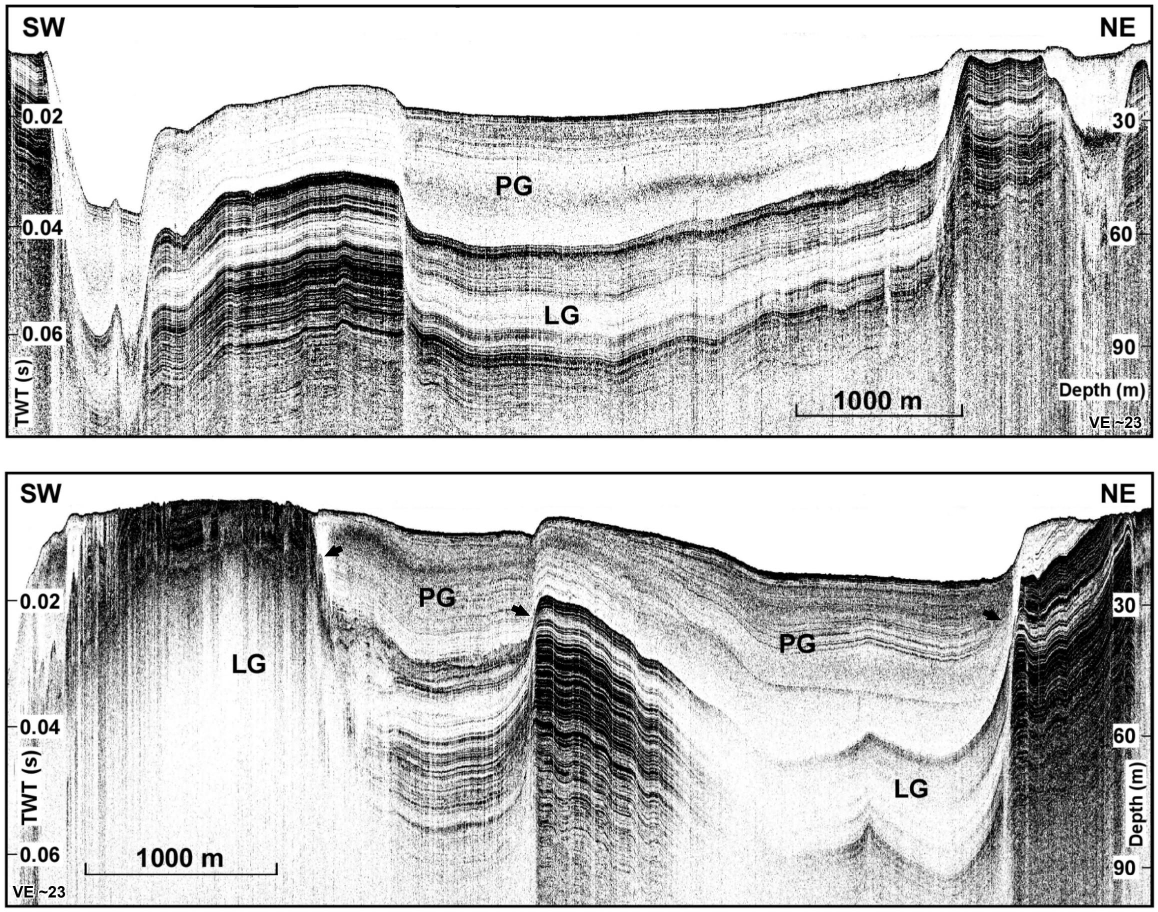

Fig. 11 (A) Seismic reflection profile 121 across Paulson Bay (Fig. 5A). (B) Seismic reflection profile 96 across Paulson Bay (Fig. 5A). Note abrupt juxtaposition of postglacial sediments (PG) against lateglacial Lake Barlow sediments (LG) resulting from faulting (arrowed).

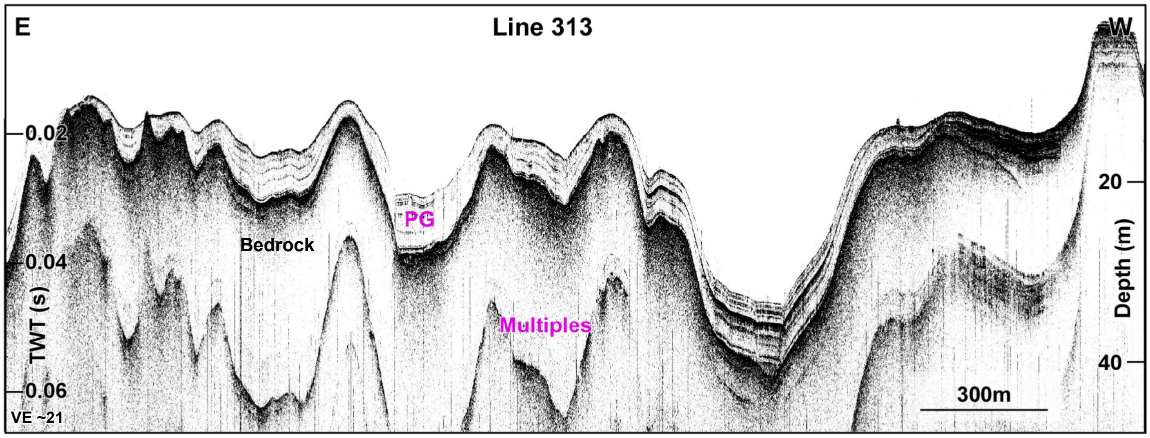

Fig. 12 Seismic reflection profile from the southernmost ‘river-lake’ part of Lake Timiskaming where postglacial sediments rest directly on crystalline bedrock of the Canadian Shield.

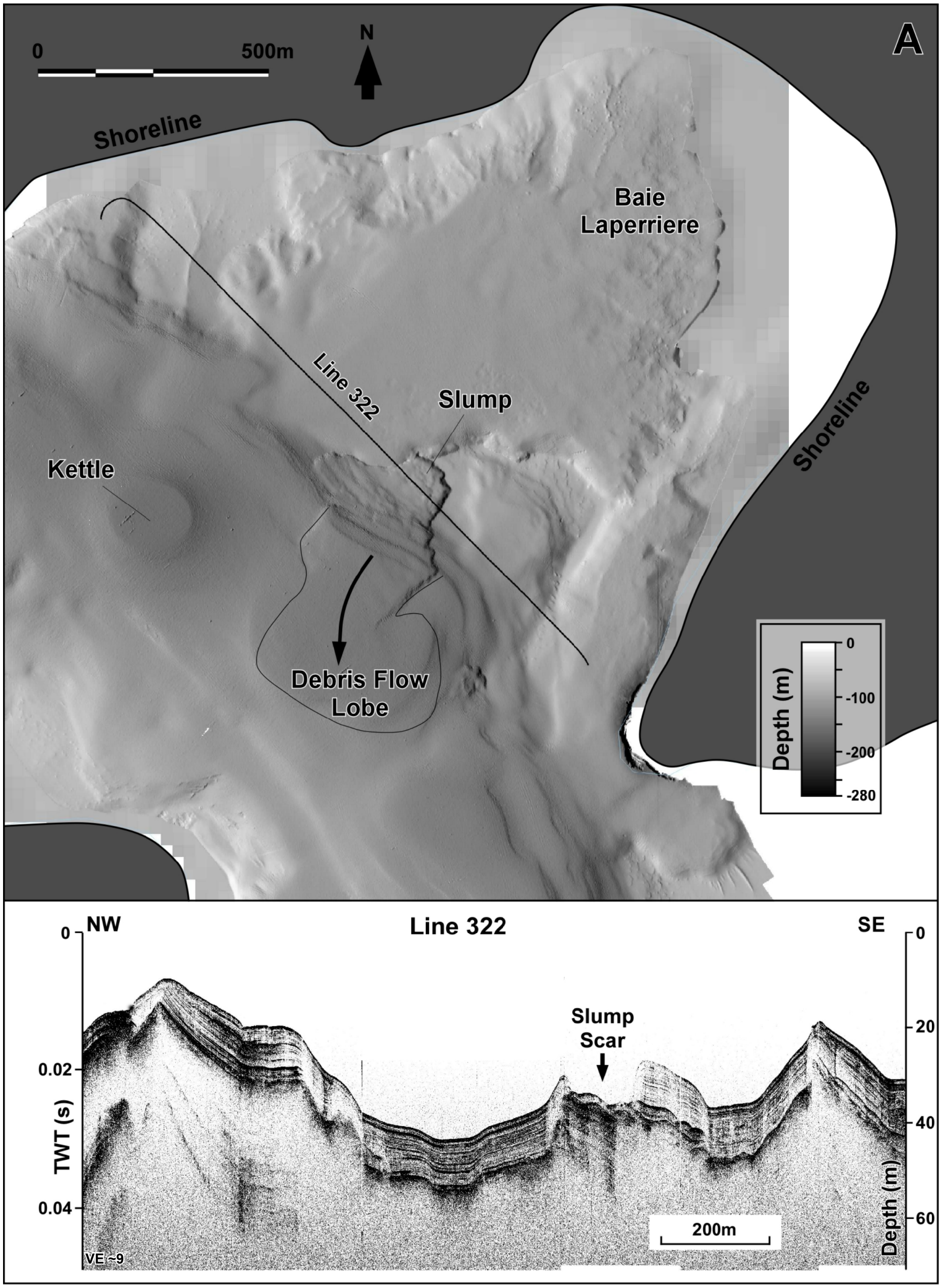

Fig. 13 (A, B) Bathymetry and seismic profiles across prominent debris flow lobes and slump scars in the central part of Lake Timiskaming. Location is shown on Fig. 5B.

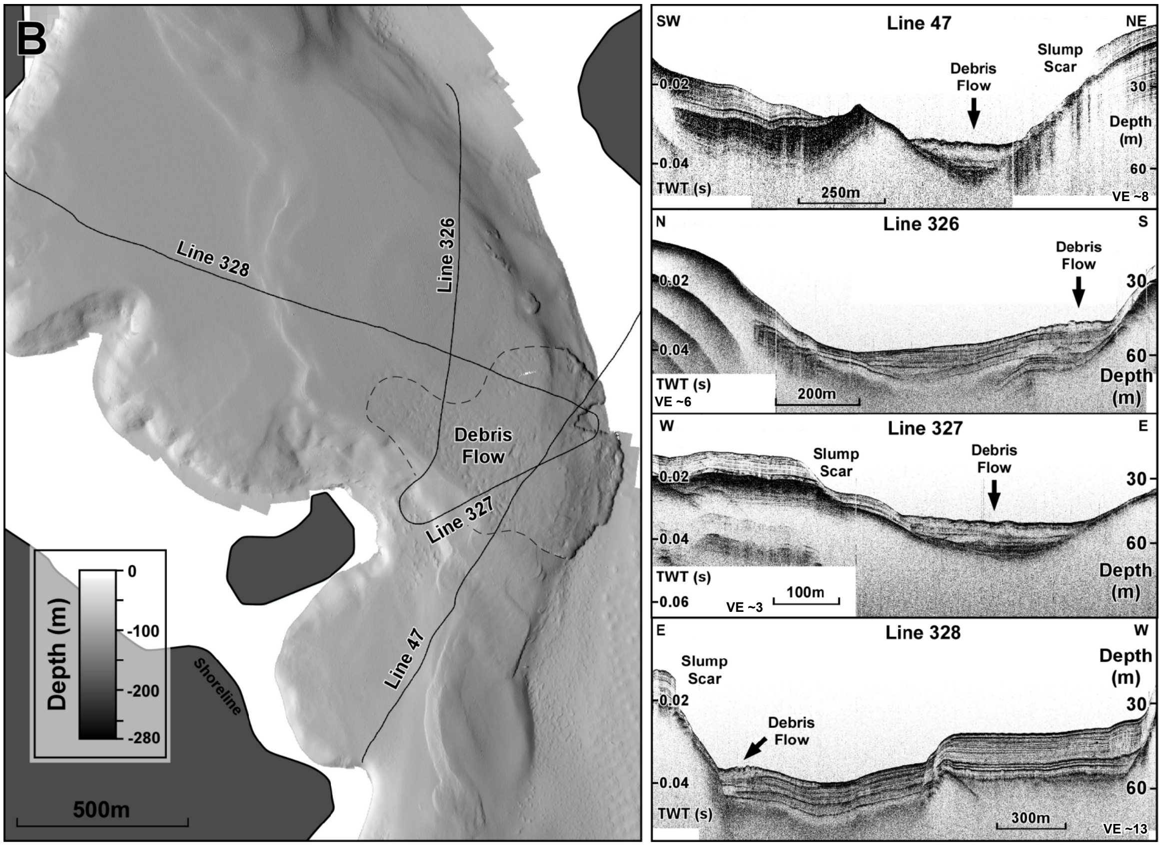

Fig. 13 Continued ...

Fig. 14 (Next Page) (A) Relationship between detailed bathymetry determined from multibeam data and underlying structures determined from seismic track lines (B) in sediment fill between Mann Island and Dawsons Point in Paulson Bay (see Fig. 5A for location). Sedimentation rates are highest (0.5 cmyr-1) in this part of the basin and the thickness of postglacial sediment approaches 40 m. Expression of faults on the lake floor suggests recent faulting given that older structures would be infilled and buried. Rectilinear fault pattern (example) arrowed.

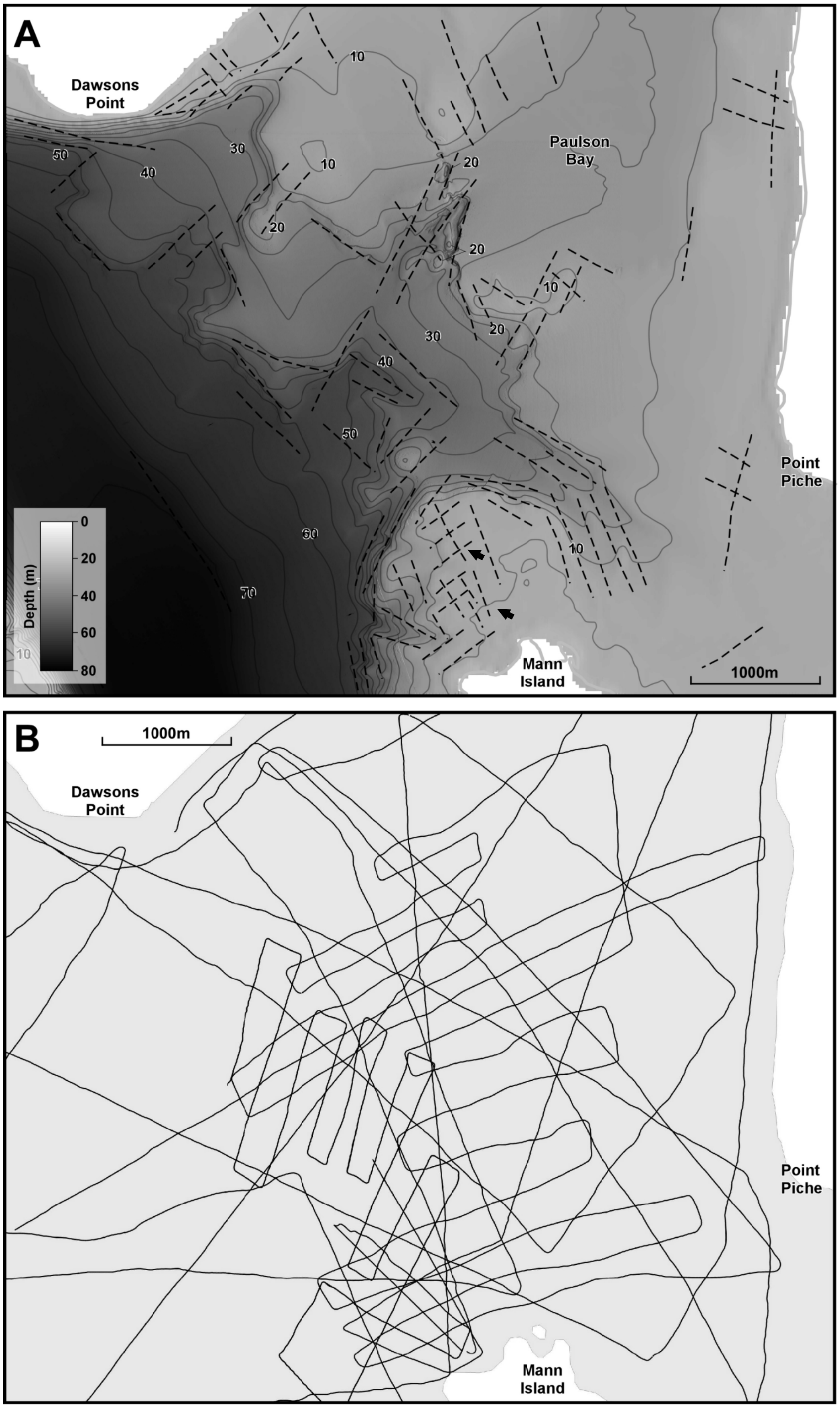

Fig. 14 Continued ...

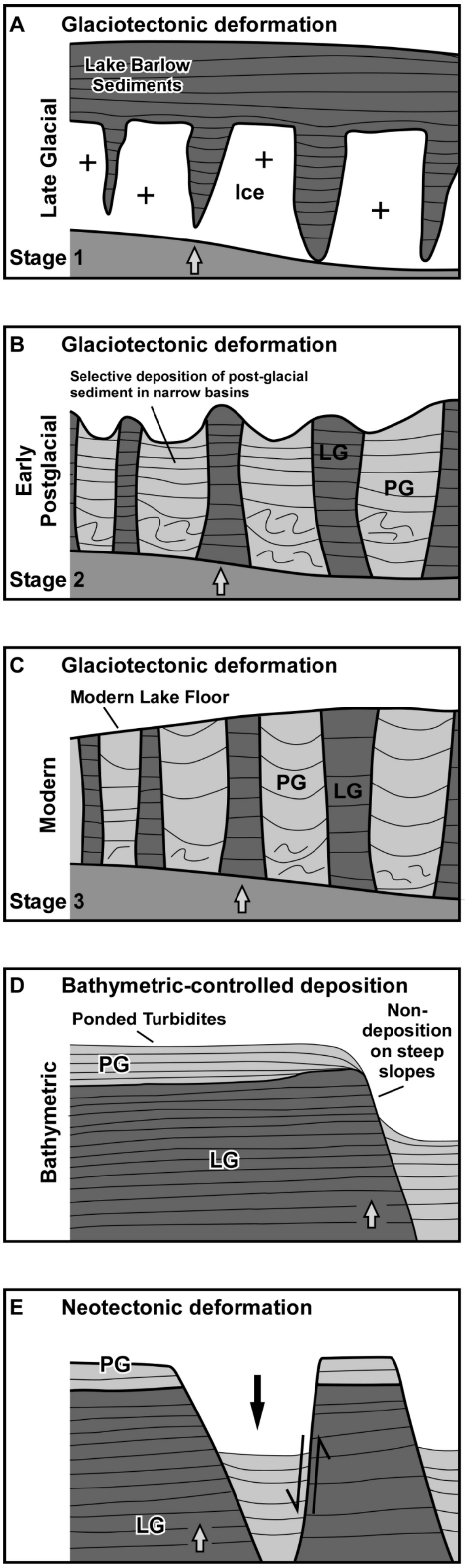

Fig. 15 Simplified models for development of closely-spaced horsts and grabens near Mann Island. (A,B,C) Glaciotectonic deformation (after Shilts, 1984). (A) Deposition of Barlow sediments on crevassed dead ice on floor of lake; (B) Active subsidence of graben basins and accomodation of early postglacial sediment in fault-contact with Barlow sediments; (C) Modern Lake floor. (D) Bathymetrically-controlled deposition. (E) Neotectonic deformation. The area of comparison (on seismic section or in plan) between models is indicated by a vertical gray arrow.

Chapter 5

Lake sediments as natural seismographs: earthquake-related deformations (seismites) in central Canadian (Ontario and Quebec) lakes

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ABSTRACT

Unconsolidated fine-grained sediments in lake basins are ‘natural seismographs’ with the potential to record ancient earthquakes. In central Canada, the lacustrine seismographic record began approximately 10,000 years ago with the retreat of the Laurentide Ice Sheet, older records having been removed by glacial erosion. Most bedrock lake basins are structurally-controlled and underlain by the same Precambrian basement structures (shear zones, terrane boundaries and other lineaments) implicated as the source of ongoing mid-plate earthquake activity. This paper highlights results of a multi-year seismic sub-bottom survey of lakes Gull, Muskoka, Joseph, Rousseau, Ontario, Wanapitei, Fairbanks, Vermilion, Nipissing, Lake Huron, Georgian Bay, Mazinaw, Simcoe, Timiskaming, Kipawa, Parry Sound and Lake of Bays, encompassing a total of more than 2000 kilometres of high-resolution track line data supplemented by multibeam and sidescan sonar survey records.

All studied basins show a consistent sub-bottom stratigraphy of relatively-thick lowermost lateglacial facies composed of interbedded semi-transparent mass flow facies (debrites, slumps) and rhythmically-laminated (varved) silty clays. Mass flows together with cratered (‘kettled’) lake floors and associated deformations reflect a dynamic ice-contact glaciolacustrine environment. Exceptionally thick mass flow successions in Lake Timiskaming along the floor of the Ottawa-Timiskaming Graben within the Western Quebec Seismic Zone, point to a higher frequency of earthquakes and slope failure during deglaciation and rapid glacio-isostatic rebound. This is also suggested by faulting, diapiric deformation and slumping of coeval lateglacial sediments in Parry Sound, Lake Muskoka and Lake Joseph, all located above prominent Precambrian terrane boundaries. Lateglacial sediments are sharply overlain by relatively-thin rhythmically-laminated and often semi-transparent postglacial silty-clay laminations. A marked unconformity between the successions records dramatic reductions in water depths under a dry climate from c. 9,000 and 8,000 ybp when the Laurentian Great Lakes were disconnected. Buried horizons of in situ tree stumps recording these low water conditions are newly identified by sub-bottom profiling in Parry Sound. The effect of postglacial neotectonics is indicated by the presence of faults, slumps and debris flows in Lake Simcoe (above a terrane boundary) and especially within lakes Timiskaming and Kipawa, recording ongoing deformation within the Western Quebec Seismic Zone along the Ottawa-Timiskaming Graben. High resolution seismo-stratigraphic data presented here supports the model that ongoing mid-plate earthquake activity is a consequence of brittle deformation of the upper crust of the North American plate. Such activity appears to have been greatest during deglaciation but continues today.

Chapter Five

5.0 Introduction

Until quite recently, the Canadian Shield and offlapping Paleozoic platforms in mid-continent North America were thought to have been tectonically stable for many millions of years and thus of low seismic risk. New seismic events occurring in unexpected locations reveal 'weak points' in the North American plate associated with active crustal deformation (Adams and Basham, 1989; Mazzotti, 2007; Tremblay et al., 2003; Wallach et al., 1998). It is now widely recognized that eastern mid-continent North America experiences many small to moderate magnitude intracratonic earthquakes that are a threat to infrastructure both onshore and along the offshore continental margin (Stein and Mazzotti, 2007). Recent large earthquakes in Canada include the Timiskaming Earthquake of 1935 (M6.2) which was felt over an area of 2.5 million km2, and the 1944 (M.5.8) Cornwall Earthquake which to date is Canada's costliest in terms of damage to infrastructure (approximately $2 million – 1944 dollars; Bent, 1996). Mohajer (1997) estimated a major (M7) earthquake can be considered 'credible' in the heavily populated southern Ontario-Quebec corridor; a recent report for the Insurance Bureau of Canada by AIR Worldwide (2013) concluded that the risk of such an earthquake is 1:500 and is estimated to create $61 billion of damage to infrastructure of which only $12 billion is currently insured (AIR Worldwide, 2013). Expected ground motions are modelled to exceed the design limits for nuclear generating facilities at Pickering and Darlington along the northern shore of Lake

Ontario near Toronto. The instrumental record is sparse for many areas of mid-continent, and AIR Worldwide (2013) also identified that current Canadian seismic risk maps on which building codes are based are outdated and do not include recent data.

Against a backdrop of increased risk, mid-plate seismicity is not yet well understood but appears to be the product of the reactivation of Precambrian structures often deeply buried below Phanerozoic strata that together record major phases of plate tectonic activity and crustal accretion and rifting along the eastern North American continental margin (Bartholomew and Van Arsdale, 2012; Cox et al., 2012; Tremblay et al., 2003 Boyce and Morris, 2002; Eyles et al., 1993). Of particular concern in eastern North America is the Western Quebec Seismic Zone (Forsyth, 1981) a belt of enhanced seismic activity that extends from Canada into the coterminous US and poses a regional risk to mines, critical nuclear facilities and the large urban areas of New York, Montreal and Ottawa (Fig. 1). In the case of the nation's capital (and Montreal), seismic risk is compounded by the presence of soft glaciomarine sediments (Leda Clays) and aged infrastructure (Motazedian and Hunter, 2008; Rosset and Chouinard, 2008). The Val-des-Bois Earthquake (M.5.0) of 2010 caused the strongest shaking ever measured in the city of Ottawa (Natural Resources Canada, 2013). Geomorphic evidence of large (M>7) earthquakes at 7060 years before present (ybp) and 4550 ybp occurs close by in the form of landslides in Leda Clays within the Ottawa Valley (Aylesworth, 2007; Aylesworth et al., 2000).

There are many practical difficulties and data gaps to be faced in fully understanding mid-plate seismicity and underlying controls on earthquakes and causal mechanisms have been said to be 'elusive' as yet (Bartholomew and Van Arsdale, 2012). The major challenge is that many Precambrian structures are not exposed at surface but are hidden below thick Paleozoic strata and Pleistocene glacial sediments. Moreover, the historic record of earthquake activity only began relatively recently in the early seventeenth century and the instrumental record is also of very short duration (typically <75 years) complicating efforts to determine recurrence intervals for moderate to large events. The lack of a detailed network of recording instruments across such a large area is another problem; regions once considered to be of low seismicity often simply reflect an absence of data (Mazzotti, 2007). Another area of debate is whether mid-plate seismicity in the glaciated portion of North America is principally related to continuing glacio-isostatic rebound and thus can be expected to decrease in the future, or reflects persistent ongoing movement of the North American plate (Ma et al., 2008). A similar debate is occurring in northwest Europe (Morner, 2011). Given the lack of any lengthy historic record in eastern North America, emphasis has been placed on examining the geologic and geomorphic record of seismic activity including the sub-bottom stratigraphy of lakes and other waterbodies (e.g., Doughty et al., 2013, 2010a; Aylsworth, 2007; Cauchon-Voyer et al., 2008; Talwani and Schaeffer, 2001; Tuttle, 2001; Aylsworth et al., 2000; Ouellet, 1997; Kelson et al., 1996; Shilts and Clague, 1992). The eastern mid-continent North America is particularly suited to this approach given the widespread distribution of lakes across the region. The present paper represents a significant contribution to a growing data set on the use of lake sediments as 'natural seismographs' in mid-continent North America (e.g., Strasser et al., 2013; Moernaut et al., 2009).

5.1 Study Area and Objectives of this Paper

This study presents the results of one of the largest surveys of lake basins conducted anywhere in eastern North America to date. It encompasses 16 lake basins across an area of approximately 100,000km^2 extending from Sudbury in the north, south to Toronto, and from Lake Huron in the west to Timiskaming and Kipawa in western Quebec adjacent to the border with Ontario (Fig. 1, 2, 3A-C). The study area includes the southernmost part of the exposed Canadian Shield which consists of Precambrian strata, and the adjacent Ontario-Erie lowlands where Precambrian basement and associated structures are buried under Lower Paleozoic platformal sedimentary strata that thicken southwards towards the Michigan and Appalachian basins. The lakes surveyed and reported in this paper were selected because of their proximity to prominent mid-Proterozoic basement structures such as terrane boundaries, lineaments and shear zones (Fig. 2, 3A).

This paper commences by briefly describing the geological evolution and resulting bedrock structure of eastern North America and its relation to mid-continent seismicity using a compilation of historic (post 1627 AD) earthquake epicentres. It then reviews and synthesizes more than 2000 line kilometres of high-resolution sub-bottom seismo-stratigraphic data from 16 lake basins with the principal objective of identifying areas of co-seismic deformations in lake sediments and their geographic location relative to underlying Precambrian structures.

5.2 Seismicity and Regional Structure in Eastern North America

A regional compilation of earthquake epicentres reveals a strong geographic relationship with the broad scale tectonic subdivision of eastern North America. Earthquake epicentres for Ontario, western Quebec and the northeastern United States for events of M2.5 or greater are shown on Figure 1. Canadian earthquake information was generated from the Geological Survey of Canada's (GSC) Canadian Earthquake Epicentre File (CEEF) and includes events recorded from 1626 through 2013. The northeastern United States earthquake epicentre and magnitude data was retrieved from the National Center for Earthquake Research (NCEER) catalogue for the central and eastern United States and includes seismic events from 1627 through 2010 (United States Geological Survey, 2010). Epicentre locations plotted prior to 1930 have an uncertainty of at least +/- 50 km (Halchuk, 2009) and those for 1970-1991 have an uncertainty of about 10 km (Stevens, 1994).

As related above, the emerging consensus is that mid-plate seismicity is related to reactivation of older Precambrian structures (Cox et al., 2012; Wallach et al., 1998; Eyles et al., 1993) and this model is supported by data presented on Figures 1 and 3 despite some uncertainty in the precise location of historic epicentres. While a comprehensive review of the geological history is not possible here, broadly parallel northeast-trending belts of epicentres identify the successive positions of the North American continental margin during the formation and breakup of the supercontinents Rodinia from 1.5 to 1.0 billion years before present (Ga ybp) and Pangea (from 450 to 350 Ma ybp) both times when large areas of crust were accreted to or detached from the margin. The Grenville Front Tectonic Zone is a wide suture zone marking the approximate continental boundary of North America at c. 1.5 Ga (Fig. 1) when South American crust (Grenville Province; GP) was added to North America during the Grenville Orogeny. The subsequent breakup of Rodinia between 750 to 600 Ma ybp is recorded by a northeast trending boundary some 200 km to the east marking the margin of the Iapetus Ocean (Iapetan Rifted Margin; Fig.1). Eastward, the Appalachian Front records the inland limit of compressional deformation associated with the building of Pangea after 440 Ma ybp when African crust was accreted to North America and in turn, this is paired with an outermost Atlantic Rifted margin formed by Mesozoic breakup of Pangea during the Late Triassic at c. 220 Ma ybp (Fig. 1).

It is very noticeable that the regional distribution of earthquake epicentres mirrors the broad structure of eastern North America outlined above, in particular the two parallel belts marking the Iapetus and Atlantic rifted margins (Fig. 1). The Iapetus margin underlies the modern St.Lawrence River and is reflected in a clustering of numerous epicentres along the so-called St.Lawrence Rift System within what is known as the Western Quebec Seismic Zone (WQSZ) (Duchesne, 2007; see below; Tremblay et al., 2003; Kumarapelli and Saull, 1966). The WQSZ shows two distinct but still poorly understood clusters (herein called 'sub-zones') that form crudely oriented northwest-trending belts separated by a zone of lower activity. The westernmost sub-zone directly underlies the Ottawa-Timiskaming Graben and lakes Timiskaming and Kipawa (Fig. 3A) extending west along the Ottawa and Timiskaming Grabens into Northern Ontario and east into the US along the Hudson Valley as far as New York City. The largest population centres in eastern North America lie within these two belts underscoring the need to understand intraplate seismicity.

The regional geology of the Canadian Shield within the study area in Ontario and Quebec consists of a complex mosaic of terranes within the Grenville Province (Fig. 2) produced by prolonged obduction during the Grenville Orogeny. The Grenville Front Tectonic Zone (GFTZ) forms a prominent structural feature under Lake Huron and emerges on land just south of Sudbury where it abuts Mesoproterozoic and Archean rocks to the north. It was active at several times in the early Paleozoic and produced seismites in carbonate and clastic sedimentary facies of the Michigan Basin (McLaughlin and Brett, 2006; Wallace, 2013a, b). The Grenville Province is subdivided into the Central Gneiss Belt (CGB) to the west and the Central Metasedimentary Belt (CMB) to the east separated by the Central Metasedimentary Belt Boundary Zone (CMBBZ) which passes southward under Pickering Nuclear Generating Station on the north shore of Lake Ontario (Fig. 2). Each belt is subdivided further into component terranes such as the Go Home, Muskoka, Bancroft, Harvey-Cardiff Arch (or domain), Belmont, Grimsthorpe, Mazinaw, Sharbot

Lake and the complex Frontenac-Adirondack Belt (Easton, 1992). To the west, in Georgian Bay, the Parry Sound Shear Zone lies along the border of the Shawanaga and Parry Sound terranes and underlies the glacially overdeepened basin now occupied by Parry Sound (Fig. 2). In their evaluation of seismic source zones Bartholomew and Van Arsdale (2012) also highlighted the additional importance of post-orogenic extensional faults formed at the conclusion of the Grenville Orogeny. These are expressed as NW-SE oriented extension fractures and faults associated with a ‘basin and range’ topography that developed across the study area after 1 Ga ybp. Such fractures may explain the strikingly linear form to the eastern shoreline of Lake Huron which is associated with the seismically active Georgian Bay Linear Zone of Wallach et al. (1998; Fig. 3A). These authors presented a detailed review of seismicity in southern Ontario and identified several seismically active basement lineaments that converge in a distinct cluster at the western end of Lake Ontario basin, the most populated area in Canada.

5.2.1 Structural controls on topography

The Canadian Shield is a low relief (< 100 m) undulating peneplain dominated by Precambrian bedrock exposed as knobs with many lakes. This surface is the product of deep differential weathering during the Mesozoic and Tertiary followed by uplift and stripping of soft weathered regolith. Final removal of weathered material was accomplished within the last 2 million years by Pleistocene ice sheets resulting in the exposure and glacial scour of fresh unweathered rock. Regolith has rarely been preserved below later glacial fills (Dyke, 2004). Weathering and glacial erosion were deepest along fractures and faults such that the Shield is geomorphologically a ‘glacially-scoured etch plain’ with a very pronounced structural control on its topography and the location of the many hundreds of lake basins on the Shield surface. Evidence of Phanerozoic basement reactivation in the study area is provided by the very close association between modern day topography cut by rivers and glaciers on top of Paleozoic strata and underlying Precambrian basement. Many lakes occupy bedrock basins that are fault controlled by underlying Paleozoic strata (Sanford, 1993; Fig. 4). Faulting occurred during a succession of Appalachian orogenies (c. 440 to 350 Ma years ago) associated with the assembly of Pangea. Seismic activity at this time triggered seismites in unconsolidated sea floor carbonate and clastic sediments in the form large ‘pillowed’ sandstones, faults and debrites that record liquefaction of sediment, loading into disturbed mud as a consequence of reverse density gradients, and mass flow activity (Wallace, 2013 a, b). Subsequently, much of the eastern part of the study area underwent uplift, unroofing Paleozoic strata and crustal stretching during the breakup of Pangea and the opening of the North Atlantic (c. 220 Ma years ago). As related above, the largest structural feature produced at this time was the St. Lawrence Rift System, a large failed rift that underlies the modern St.Lawrence River valley (Cauchon-Voyer et al., 2008; Kumarapelli and Saull, 1966; Fig. 1). In addition to a close structural control on the location of lake basins, the regional drainage patterns even in areas of thick glacial cover are controlled by neotectonic jointing which affects all strata regardless of age. Maximum regional horizontal stress is consistently clustered around a NNE to E direction and regional mapping show a consistent grouping of joint orientations into NE-SW and NW-SE trending sets in both Paleozoic and Precambrian rocks (see Lam et al., 2007; Eyles and Scheidegger, 1999, 1995; Andjelkovic et al., 1998; Eyles et al., 1997).

5.3 Glacial History of Study Area

The study area was glaciated many times during the Pleistocene but a lacustrine sedimentary record only exists for the final phases of the last (Wisconsin) glaciation. The area was entirely covered by the Laurentide Ice Sheet which reached its maximum thickness and extent in mid-continent at about 20,000 ybp (Dyke, 2004). Lacustrine sediments containing warm-to-cool climate indicators spanning the last interglacial through mid-Wisconsin glaciation are preserved in buried bedrock valleys at Toronto but to date these have not been identified in any modern lake basin and are presumed to have been eroded elsewhere. Lake basins were apparently cleaned out to bedrock by glacial erosion and backfilled during deglaciation.

The Shield is, in general, lacking in sediment cover and exposed glacially-scoured bedrock predominates with patches of lake sediments left by large deglacial lakes (see below). Glacial sediments up to 100 m thick, thicken southwards onto Paleozoic strata and bedrock is only exposed in the deeper valleys. Deglaciation began in the south about 12,000 ybp and cleared the most northern parts of the study area at Timiskaming by about 9,000 ybp (Fig. 5). Very large high-level lakes formed against the retreating ice margin such as glacial Lake Algonquin in the Huron basin, glacial Lake Iroquois in the Ontario basin and Barlow-Ojibway along the length of the Timiskaming Graben (Jackson et al., 2000). Lake levels fell abruptly in the early postglacial as water drained through the ice free but still glacio-isostatically depressed Ottawa and St. Lawrence valleys (Fig. 5). There is emerging evidence that the Laurentian Great Lakes were disconnected shortly thereafter and lake levels were lowered by as much as 60 m below modern levels as consequence of a warm dry climate in mid-continent just after 9,000 ybp (McCarthy and McAndrews, 2012). Large areas of the floors of the modern lakes Erie, Huron and Ontario were exposed at this time recorded by brackish closed basin conditions, and greatly diminished outflows along the Niagara River. Lake levels have risen subsequently, as recorded by drowned beach systems and flooded archaeological sites, as a consequence of cooler and wetter climates and regional tilting associated with enhanced glacio-isostatic uplift of the outlet of the Great Lakes at the eastern end of Lake Ontario (see Jacobi et al., 2007 and Jackson et al., 2000 and for additional details).

5.4 Geophysical Methods

The high resolution ‘chirp’ seismic reflection survey data described here was collected using an EdgeTech high resolution XStar or 3200-XS digital sub-bottom profiling system (see EdgeTech, 2009 and 1998) with a SB 216S tow vehicle (‘fish’). The system transmits an FM sonar ‘chirp’ pulse created by linear sweeping over the frequency range of 2-12 kHz for 20 ms (see Eyles et al. 2003). Sediment and water depths displayed on the seismic stratigraphic profiles assume a constant velocity of 1450 m/s (e.g., Doughty et al., 2013, 2010a; Cauchon-Voyer et al., 2008; Strasser et al., 2007; Mullins and Eyles, 1996). Water depth data determined from seismic data agree with charting by the Canadian Hydrographic Survey. Track line density is sufficient in some cases (Parry Sound, Lake of Bays, Timiskaming and Joseph) to allow seismo-acoustic fence diagrams to be constructed. Vehicle positioning was recorded from standard Garmin-GPS instruments. XStar data were streamed to either external JAZZ-type removable hard disks or to DAT, both in XStar SEGY format (a modification of the standard SEGY format; Barry et al., 1975; EdgeTech, 1998). After transferral to a separate workstation, data processing steps included: translation to standard SEGY format; partitioning into appropriate track-lines/lengths; conversion of coordinates from minutes-of-arc (i.e. latitude/longitude) to UTM; correction of coordinates due to GPS tracking errors; conversion of coordinates from points-to-line for incorporation into a GIS; conversion into Seismic Unix SEGY format (Cohen and Stockwell, 2012) for visualization and any further processing. In addition, the envelop amplitude was calculated for each seismic/track-line resulting, generally, in improved imagery. Most conversions, when error-free, occurred within a batch framework (consisting of various conversion scripts). 3200-XS data are collected and stored directly by the field unit as EdgeTech JSF formatted files (EdgeTech, 2010). Similar data processing steps (to the XStar, above) were applied with the addition of a JSF-to-SEGY conversion. Seismic/sub-bottom imagery was also derived directly using the standard 3200-XS software package.

5.5 Lacustrine facies types and successions

Seismic profiling of sediments preserved on lake floors using the XStar system provides excellent high-resolution seismo-stratigraphic data typically to a maximum sub-bottom depth of ~ 60 m in fine-grained silty clay sediments (Figs. 6 - 22). This depth window allows full penetration of the entire thickness of the sedimentary fill of most lakes down to bedrock with the exception of Lake Timiskaming where the base of the Pleistocene sediments is too deep to be resolved. Acoustic penetration in more coarse-grained lake floors (i.e. sand, bedrock) is typically less than 5 m.

Most of the lakes investigated are characterised by an undulating bedrock basement of moderate to high relief, often sufficient to protrude through the entire sediment cover and isolate sediments into sub basins separated by bedrock highs devoid of any sediment. Bedrock forms a hard 'acoustic basement' that is evident on many seismic profiles. Notably, no primary glacial sediments such as till or glaciofluvial outwash gravels or any glacial landforms such as moraine ridges have been observed on seismic profiles. This confirms the model that lake basins were essentially swept clean by ice and that deglaciation was rapid and not punctuated by still stands and deposition of coarse grained primary glacial sediment. Instead, bedrock basins only became active depocentres for relatively fine sediment as the ice margin withdrew in contact with deep lateglacial lakes.

5.5.1 Lake sub-bottom stratigraphy

Two superposed acoustic stratigraphic successions can be identified in lakes across the study area forming a consistent bipartite bottom stratigraphy (Figs. 6 - 22). The presence of a stratigraphic couplet that records lateglacial and postglacial phases of sedimentation was first recognized by Shilts (1984) and was confirmed by subsequent work (e.g., Lazorek et al., 2006; Eyles et al., 2003; Moore et al., 1994; Rea et al., 1994; Shilts and Clague, 1992, and references therein). These two successions are expressed on seismic profiles as a lowermost, darker-coloured, more reflective succession resting on bedrock (designated LG for late glacial; e.g. Fig. 6) and characterized by closely-spaced, high-frequency parallel reflectors typical of rhythmically-laminated glaciolacustrine silty-clays deposited during deglaciation in proglacial (ice-contact) lakes. In contrast, an upper thinner (typically less than 5 m) and largely semi-transparent seismic succession (designated PG for post glacial; Figs. 6, 7, 8) shows much weaker acoustic returns and very closely-spaced parallel reflectors typical of modern annually-deposited postglacial sediments dominated by organic material, typically silica-rich gyttja derived from diatoms (Lazorek et al., 2006; Shilts and Clague, 1992). Weaker acoustic returns in the postglacial succession are the consequence of much reduced rates of clastic minerogenic sedimentation compared to those during deglaciation; many lake basins are surrounded by bare rock devoid of sediment are consequently ‘sediment starved.’ Exceptionally, the postglacial succession in the northern part of Lake Timiskaming is on average between 15-20 m thick and up to 40 m adjacent to the mouths of the large rivers that rework exposed lateglacial lake sediments in the surrounding basin. Modern sedimentation rates in Lake Timiskaming are as high as 0.5 cm/yr-1 (Doughty et al., 2013; Shilts, 1984).

While the existence of a bipartite acoustic sedimentary succession in glaciated Shield lakes in Canada is well established, new significance is now placed on the climatic significance of the unconformity that commonly separates them. This erosion surface truncates underlying lateglacial facies (e.g. Fig. 9) and is the consequence of abrupt falls in lake levels across the region when large glacial lakes impounded against the retreating Laurentide Ice Sheet suddenly drained (Fig. 5). Coring and dating of the unconformity in lakes Nipissing and Huron suggests that the unconformity also records a lengthy 1000 year-long episode of dry climate shortly after deglaciation between c. 9,000 and 8,000 ybp when runoff was drastically reduced throughout the Great Lakes basin and closed basin conditions obtained (McCarthy et al., 2012; Lewis et al., 2007). A dramatic expression of this event can be seen on seismic profiles of the stratigraphy below the floors of several protected inlets along the northern margin of Parry Sound (Fig. 3C, 9A). There the unconformity is well expressed as a dark, highly reflective zone some 5 m thick at the base of overlying Holocene sediments. This horizon occurs at a consistent elevation of some 50 m below modern lake levels and is characterised by numerous closely-spaced point source reflectors producing a distinctive ‘spotty’ seismofacies type that has not been seen in other lake basins. These seismofacies are typically associated with ice-rafted debris in fine lacustrine sediment but the size and number of point reflectors is atypical and forms extensive horizons. Instead, these facies are tentatively interpreted as the rooted stumps of sub fossil trees buried by lacustrine sediment during an increase in lake levels as climate cooled after 8,000 ypb.

5.5.2 Lateglacial debris flow

This study highlights the important role of debris flow during lateglacial sedimentation, hitherto unrecognized in existing models of glaciolacustrine sedimentation (Ashley, 1975). An important sedimentary process operating during lateglacial time in Canadian lake basins has been seasonally-active, density-driven underflows (quasi-continuous turbulent flows) and overflows resulting in accumulation of high-frequency rhythmically-laminated facies. These are likely of annual origin ('varves') but this cannot be positively demonstrated in the absence of cores from lake bottom sediments. However, Antevs (1925) studied exposed outcrops of lateglacial lacustrine sediments in the northern part of the study area and demonstrated an annual control on sedimentation. Of considerable interest here is the common presence in lateglacial successions of thickly-bedded 'chaotic-massive-to-stratified' beds ('c-m-s facies'). These are seen as distinct transparent to weakly reflective intervals within lateglacial successions of high-frequency, rhythmically-laminated glaciolacustrine silty-clay (Figs. 6, 8, 9). 'C-m-s' facies have undulating lower bounding surfaces which are conformable or erosional. In the latter case, these units are chaotic consisting of a mix of darker and lighter coloured zones interpreted as rafts of underlying laminated facies. Upper surfaces are frequently hummocky in form. Chaotic seismic facies pass laterally and vertically into massive transparent facies that in turn, are overlain by high-frequency rhythmically-laminated facies marking a return to rhythmic sedimentation (Figs. 8, 9). The upper surface of c-m-s beds is frequently hummocky and beds thicken downslope. 'C-m-s' beds are readily interpreted as sediment gravity flow facies, specifically those formed by debris flow (debrites) based on their description from cores in other basins (Strasser et al., 2011; Locat and Lee, 2002). Basal chaotic facies within erosionally-based 'c-m-s' beds are interpreted as relatively proximal facies resulting from slumping and incomplete downslope mixing of masses of semi-coherent unconsolidated rhythmically-laminated facies. A hummocky surface to such flows is consistent with downslope compressional thickening during flow (e.g., Joanne et al., 2013). Water escape structures (seen as narrow vertical pillars; e.g. Fig. 8) are very common indicating the development of overpressured conditions in such deposits

Massive facies (mainly transparent signatures) are interpreted as the more completely homogenised ('fully mixed') and further-travelled portion of debris flows. The same seismo-facies were identified on high resolution seismic records from Seneca Lake, the largest and deepest of the New York Finger Lakes by Halfman and Herrick (1998). Limited outcrop work to date has recognized downslope mass flow deposits in the form of slumps and massive, apparently disorganized sand beds (see Kelly and Martini, 1986; Kaszycki, 1985). This study shows that massive facies are a volumetrically significant part of the lateglacial fill of lake basins. Hitherto, the extent of mass flow activity in lake basins during deglaciation has not been recognised (see Shilts et al., 1992) largely because of the relatively limited seismic data available (Shilts and Clague, 1992).

Unfortunately modern outcrops through lateglacial deposits containing mass flow facies are rare as these are usually covered by slumped colluvium. Figure 10 shows disturbed near-surface horizons of lateglacial glacial Lake Barlow-Ojibway silty-clay laminations briefly exposed by road construction near Ville Marie on the eastern side of Lake Timiskaming in Quebec. This outcrop is considered significant because Antevs (1925, p. 81) presented a detailed description of outcrops of the same lateglacial age in the nearby Ville Marie-Bearn area that contain a very prominent intraformational deformation horizon. This regionally correlative horizon was described as being ‘many feet’ thick, faulted, tilted, and contorted with intraformational rafts and blocks of undisturbed ‘varves’ lying stratigraphically directly above his ‘varve 498’. This deformed horizon fulfills one of the most important criteria for recognition of co-seismic origin of soft sediment deformation, namely widespread distribution. It can be suggested that Antevs (1925) was reporting a thick debrite bed produced by co-seismic slumping similar to that seen on sub-bottom seismic profiles of Barlow-Ojibway deposits below the floor of Lake Timiskaming (Figs. 6, 7). The presence in seismic data of what are interpreted as large rafts of laminated silty clays in these facies (Fig. 8) agrees with the outcrop observations of Antevs (1925). Unfortunately, the Barlow-Ojibway deposits are too extensively disturbed by faulting below the floor of Lake Timiskaming (see Doughty et al., 2013) to allow meaningful determination of the stratigraphic frequency and down basin extent of such beds. Nonetheless, they are a very common component of the basin fill and are thick and extensive indicating repeated large scale mass flow events during deglaciation. In this regard, it is highly significant that examination of Antevs (1925) detailed descriptions of many other outcrops throughout the Timiskaming district (pp. 105-115; Antevs, 1925) reveals the common presence of faults and slumped horizons within ‘varved’ deposits (which he often attributed to glacial activity such as iceberg scour) together with unusually thick (1.5 m) beds he attributed to sudden inputs of meltwater (the so-called ‘drainage varves’ of Antevs, 1925). Given the seismicity of the Timiskaming Graben these structures are suggestive of repeated co-seismic deformation involving faulting and debris flow. Adams (1989, 1982) suggested that a record of earthquake activity is preserved as deformed horizons in the laminated successions of the Timiskaming basin; geophysical data presented here, together with the previous observations of Antevs (1925), provide strong support for this hypothesis and suggest that further work is warranted. Debrites are very common in other lakes of the study area (Fig. 8, 11) but it is significant that the thickest and most spatially extensive occur below Lake Timiskaming within the seismically-active Western Quebec Seismic Zone.

Subaqueous debris flows are triggered by many processes in dynamic ice-contact glaciolacustrine environments such as the melt of buried ice (see below), ice berg calving and scouring, meltwater flood events, sudden lake drainage or depositional oversteepening, and of course by earthquakes. The debris flows found in the lateglacial record of Lake Timiskaming, within sediments deposited in glacial Lake Barlow-Ojibway (Fig. 6, 7), are unusually thick (~ 5 m) and extensive (2 km) suggesting earthquake-triggered slumping akin to the modern slumps seen along the lake's margins and in nearby Lake Kipawa (see below; Doughty et al., 2013). This appears likely, but cannot at this stage, be demonstrated. Moernaut et al (2007) and Strasser et al (2013, 2007) describe similar earthquake-triggered mass flows from other large lakes comparable in thickness and extent to those described here.

5.5.3 Lateglacial faulting and diapirism related to possible earthquake activity

In addition to subaqueous slumping and mass flow triggered by seismic activity, the present study identifies several other examples of lateglacial deformations (faults and diapirs) that are clearly unrelated to the melt of buried ice or slumping on the floors of Parry Sound, Lake Muskoka and Lake Joseph. These structures lie directly above Precambrian terrane boundaries (Figs. 8, 11, 12) and can be convincingly related to seismic activity during deglaciation. The structures are restricted to lateglacial sediments only and are truncated by the unconformity forming the base of the overlying postglacial acoustic succession (Figs. 8, 9). This provides a precise constraint on the timing of faulting during the brief (500 year) lifetime of glacial Lake Algonquin that formed along the ice front during deglaciation and suggests that faulting is a consequence of rapid postglacial rebound immediately after deglaciation.

Faulting of lateglacial sediment, the vertical escape of gas/water and large-scale diapiric movement of water-saturated sediment is seen below a thin cover of postglacial sediments where Lake Joseph empties into Lake Rosseau (Fig. 12B). Step-like extensional faulting occurs over some 2 km length of the lake floor giving rise to marked offsets of correlative horizons within the lateglacial stratigraphy (Fig. 9C). These structures are associated at depth on seismic records by narrow tapered 'pillars' of massive unstructured facies typically produced by upward escape of fluid or gas (Fig. 9C). Larger scale upward movement of sediment toward the lake floor is also seen in the form of a diapiric 'mus-hroom-like' structure (Figs. 12A, 12B). A thin debris flow with a distinctly hummocky surface extends downslope toward this structure. Diapirs associated with pits and linear troughs on the lake floor and produced by erosion of sediments by escaping fluids or gas, were also noted above terrane boundaries in western Lake Simcoe by Boyce et al. (2002).

5.5.4 Lateglacial deformation related to melt of buried ice

Many of the studied lakes contain deformed sediments resulting from the melt of buried glacier ice trapped along lake floors. Several lake basins, especially Lake of Bays, Gull, Mazinaw and part of Timiskaming, show areas of their floors that are dimpled ('kettled' or cratered) as a consequence of the melt of large remnant ice blocks buried under a cover of lateglacial sediment. Ice melt resulted in the collapse of overlying and adjacent sediment and the production of a crater (Fig. 13); typically associated with deglaciation of a strongly undulating often high relief bedrock surface. Large remnant ice blocks were severed from the downwasting and retreating margin of the Laurentide Ice sheet by bedrock highs protruding through the ice. This ice was trapped in confined topographically lowermost parts of narrow lake basins and weighted down by lacustrine sediment (Fig. 14; see Eyles et al., 2003). Some ice blocks may have originated as stranded icebergs. These craters ('k-ettle basins') are characterised on acoustic profiles by rings of faulted and chaotically-bedded sediment where any original stratigraphy has been disrupted by collapse and slumping. (e.g., Eyles et al., 2003; Kaszycki, 1987; Klassen and Shilts, 1982)

The rate of postglacial sedimentation has generally been insufficient (with the exception of Lake Timiskaming) to entirely fill these collapsed sub-basins which are still expressed as circular depressions in the bathymetry of the lakes. Kaszycki (1985) described the effects of thawing buried ice on glaciolacustrine sediments exposed in out-crop within the Gull River watershed, including Gull Lake itself, and suggested that all major lakes in that watershed are floored by intensely faulted laminated sand, silt and clay as a result of ice trapping and melt.

5.5.5 Postglacial faulting and slumping

Seismic data from many of the lake basins investigated in this study are also interpreted to show evidence of postglacial (Holocene) faulting and slumping that may relate to earth-quake activity. Detailed seismic data from the bottom stratigraphy of Lake Timiskaming has already been presented (Doughty et al., 2013) and a single illustration is presented here of faults that affect the entire lateglacial and postglacial Holocene fill of this lake (Fig. 7). Open fractures on the lake floor and numerous slumps around the basin margin in deep water (Figs. 15, 16) also testify to ongoing deformation related to frequent earthquakes within the Western Quebec Seismic Zone (see below).

Recent slumps on the floor of Lake Kipawa were mapped by Doughty et al (2010a) and their geographic distribution was shown to be systematically related to the epicentral location of the 1935 M6.2 Timiskaming Earthquake. Figure 15 shows a map of slumps and other lake floor features in Lake Timiskaming compiled from multibeam, acoustic and side scan datasets. The Lake Timiskaming basin consists of a long linear basin with a flat marginal ‘shelf’ of varying width in shallow water (0 to 15m depth). The shelf has a sharply-defined outer edge that overlooks a steep scarp-like slope that descends to deeper water (generally, 140 to 160m depth; a maximum of 198m has been recorded). The shelf results from planation of lateglacial glaciolacustrine sediments by wave-induced current activity in shallow water. Shelf width is greatest on the eastern side of the basin (varying between 600 and approximately 2000m at the north end of the lake, Fig. 15A) and narrowest or non-existent on the west along the exposed bedrock trace of the West Shore Fault (varying between 0 and a maximum of 600m; Fig. 15A).

The eastern slope is distinctly ‘crenulated’ with numerous semi-enclosed bowl-shaped excavations and narrow gullies along the shelf edge. Debris flow lobes are common on the lower slope and in many cases extend from the mouths of upslope bowl-shaped valleys (Figs. 15A, 15B). This suggests the importance of mass flow in the headward erosion of such valleys; indeed their abrupt headward termination is similar to the so-called 'theatre-headed' canyons of semi-arid areas where back wall retreat is accomplished as a consequence of spring sapping, undercutting of back walls and their collapse. Spring sapping as a major causative agent is likely for the crenulated form of the shelf break in Lake Timiskaming and is suggested by the form of a large 'A'-shape depression present on the outer shelf edge in Paulson Bay (Fig. 17; see also Shilts, 1984 and Shilts et al., 1992, for description of gullying). This feature consists of two narrow linear gullies that join together at their headward ends leaving a remnant high-standing block of sediment between; small enclosed depressions to the south link the two gullies (forming a crossbar as in the letter 'A') and separate another remnant block to the south. This area of the lake floor is crossed by two sub bottom track lines and the location of gullies can be seen to be controlled by faults and diapirs in underlying lateglacial glaciolacustrine sediments (Fig. 18). Subsidence and sediment collapse over growing faults is suggested by a small enclosed depression to the north. Linear open fissures are clearly observed on nearby multi-beam and side scan profiles (Fig. 19). This suggests an actively-developing lake floor geomorphology where collapse of sediment is occurring above linear fractures or faults that control fluid migration through lacustrine sediments and subsequent seepage zones on the lake floor.

The data presented above seem to indicate that groundwater seepage and spring-sapping of soft glaciolacustrine sediment is a widespread subaqueous process operating along the outer shelf edge and slope of the Lake Timiskaming basin, at least at its northern end (Fig. 15A; Kotilainen and Hutri, 2004; Jensen et al., 2002; LaFleur, 1999; Shilts and

Clague, 1992). Soft overpressured sediment is prone to mass wasting and explains the common presence of debris flow lobes on the lower slopes of bowl shaped depressions. Overpressuring of sediment and upward movement of saturated postglacial sediment is indicated by the diapiric structures that immediately underlie gullies in Paulson Bay (Fig. 18). Given the known seismicity of the area and presence of earthquake-triggered slumps in nearby Lake Kipawa (Doughty et al., 2010a) it is reasonable to argue that many debris flows have been triggered by coseismic ground shaking and slumping of glaciolacustrine sediment softened by groundwater seepage (Shilts, 1984). Doig (1991) reported deformed silty clay sediments in short (< 1m) gravity cores collected from a single site in northern Lake Timiskaming that were interpreted as the product of earthquake-triggered mass flow during the 1935 (M6.2) Timiskaming Earthquake. The present study, which focuses on basin-wide seismic data, reveals the much broader geographic extent of debris flows on the lake floor. Debris flows are as much as 5 m thick on acoustic profiles and the largest lobe covers some 400,000 m2 of the lake floor (Fig. 16). It is highly significant that debris flows are mainly present on the lake floor north of the Montreal River Fault which is the southernmost active fault in the Timiskaming Basin. Thick lateglacial sediments are present south of this fault but evidence of faulting or mass flow activity on the lake floor is lacking (Doughty et al., 2013). This suggests the southern part of the Timiskaming basin is structurally stable. Furthermore, the distinct preferential geographic distribution of debris flows on the east side of the lake as a whole (Fig. 15) suggests some directionality in seismic wave propagation or may simply indicate that the Timiskaming East Shore Fault (TESF) is more active than its western counterpart. This latter suggestion is strongly supported by geomorphic evidence seen on the graben floor north of Lake Timiskaming. Between New Liskeard and Earlton a 20 km long and strikingly linear 8m high bluff has been interpreted as the consequence of postglacial faulting along TESF and displacement of lateglacial deposits of lake Barlow-Ojibway (Doughty et al., 2013, 2010b).

Figure 20A shows the preliminary results of a lake-based magnetic survey of Kempenfelt Bay in Lake Simcoe (Fig. 2). Magnetic mapping reported briefly by Boyce et al. (2002) identifies a number of west-east trending curvilinear anomalies and a broad zone (approximately 3 km in width) of high amplitude northwest-trending magnetic anomalies that crosses the middle section of the bay. The high-amplitude anomalies mark the boundaries of Alliston-Go Home basement shear zone in the underlying Precambrian basement and reflect the presence of mylonitic rocks along the shear zone that have much higher susceptibility than the surrounding basement gneisses. The magnetic image also shows clearly that the shear zone is offset by a west-east magnetic lineament. The lineament is interpreted as the trace of a strike-slip basement fault that offsets the boundary with a left-lateral displacement of about 500 m (Fig. 20).

Side-scan sonar imaging of Kempenfelt Bay shows that the southeastern boundary of the faulted shear is associated with slumping of Holocene sediments on the adjacent basin slope (Fig. 21). The discordant outcrop patterns imaged on the slope indicate the presence of several large, rotated slump blocks. The slump blocks appear to be part of a larger, lobate mass flow deposit that extends onto the basin floor. The debris flow lobe is expressed on bathymetric data as a distinct bulge in bathymetry contours at the southeastern edge of the shear zone (Fig. 20B). Large-scale slump features were not identified elsewhere along the basin margins. The features are interpreted as evidence for Holocene slumping triggered by neotectonic reactivation of underlying basement structures (see also Boyce et al., 2002; Boyce and Morris, 2002).

Sub-bottom sonar profiles from Kempenfelt Bay show that west-east basement magnetic trends are associated with zones of normal faulting and also localized thrust faulting of the postglacial lacustrine infill (Figs. 21-22). Normal faults comprise graben-like block faults that strike parallel or sub-parallel to the basin axis (Fig. 21A); these were noted by Todd and Lewis (1993). Reverse faults occur as small 'en echelon' sets of west-east striking thrusts that show small vertical displacements (< 2 m) (Fig. 22B). Other features captured on sub-bottom profiles include gas escape structures that appear as characteristic convex-upward 'bee-hive' reflection patterns (Fig. 22A). The structures are associated with localized depressions and linear troughs in the lake floor that are produced by erosion of sediments by the escaping gas. The linearity of the trough features suggests that gases and pore waters are likely moving vertically along discrete fractures or faults in lake sediments (Boyce et al., 2002).

5.6 Discussion

This study is based on a partial survey of the many hundreds of lakes in Ontario and western Quebec but it is an important first step in a broad regional assessment of seismic risk using lacustrine records. It confirms the utility of using such sediments as natural seismographs and shows that the lacustrine fill of several water bodies (Parry Sound and lakes Timiskaming, Joseph, Muskoka) records deformation interpreted to relate to ground shaking during lateglacial earthquakes; all of these basins are structurally-controlled and lie directly above major Precambrian basement structures (Fig. 23). These deformations (faults, slumps, debris flows, water escape structures) can be collectively referred to as seismites. Those in Parry Sound (Fig. 9A), Lake Joseph (Fig. 8B) and Lake Muskoka (Fig. 8C, 11) are of lateglacial age (they are truncated by an early postglacial unconformity) and suggest the occurrence of moderate to large earthquakes during deglaciation and rapid crustal rebound. Those deformations in the Muskoka lakes occur above Precambrian terrane boundaries in the Canadian Shield directly along the margin of the Go Home and Muskoka terranes; those deformations in the Parry Sound area lie very close to the intersection of the Georgian Bay Linear Zone and Parry Sound Shear Zone (see Wallach et al., 1998; Figs. 2, 20).

Deformations identified in Lake Simcoe, Lake Timiskaming and Kipawa are, in contrast, of postglacial age as they affect both lateglacial and postglacial sediments. Those at Timiskaming and Kipawa are readily related to ongoing seismicity within the WQSZ associated with known measurable crustal deformation within the North American plate along the Timiskaming Graben (see Mazzotti, 2007). Postglacial deformations in Lake Simcoe are likely related to activity along terrane boundaries that pass southwards along the western margin of the lake and control the location of that margin (Figs. 2, 20). In Lake Ontario, Thomas et al (1993) identified postglacial bedrock 'pop-ups' in the form of WNW trending elongate features projecting 1.5 to 2m above the lake floor from side-scan sonar data (see also Jacobi et al., 2007). In addition, so-called 'plumose structures', referring to a feather-like arrangement of small-scale ridges no more than 10-15cm in height, were mapped in postglacial sediment with a distinct ENE trend. Thomas et al (1993) considered these to be analogous to similar structures reported proximal to off-shore 'pockmarks' on the eastern Canadian continental margin associated with active faults (e.g. Fader, 1991; Pecore and Fader, 1990). Thomas et al. (1993) also identified steeply-dipping scarps on the floor of southeastern Lake Ontario associated with apparent offsets of reflectors in glaciolacustrine clays. These were interpreted as normal faults with displacements of 10-15m but these so-called 'offsets' more likely reflect non-deposition of sediment on very steep bedrock slopes ('bathymetric-controlled deposition'; see Doughty et al., 2013, p. 999).

Much discussion of earthquake risk has focussed hitherto, on the seismically-active and regionally-prominent lithotectonic boundary between the Central Gneiss Belt and Central Metamorphic Belt (CMBBZ; Fig. 2) which passes southward below the Greater Toronto Area (population 5 million) and directly below the nuclear generating station at Pickering on the north shore of Lake Ontario. Faulted glacial sediments within 7 km of the plant were reported by Mohajer et al. (1992) but their origin remains elusive (see Eyles and Mohajer 2003; Godin et al., 2002) although some are considered to be neotectonic (Godin et al., 2002, p. 1389). Gull Lake lies directly above this structure to the north and although faulting is seen on acoustic profiles of lateglacial sediments, these are associated with craters (kettles) are therefore, interpreted as a consequence of the melt of buried ice. Further work will focus on other lakes along different parts of the CMBBZ to more closely resolve the Holocene history of this complex boundary zone.

The introduction to this paper highlighted the challenge of discriminating earth-quake-related deformations from non-seismic deformations produced by mass flow or glacial activity such as collapse over buried ice or glaciotectonic deformation. A major finding is that Shield lakes were cleared of any pre-existing sediment by the Laurentide Ice Sheet and only filled during deglaciation. As a consequence, deformation structures resulting from glaciotectonic overriding of older sediments which complicates identification of co-seismic deformations, are absent from these basins unlike those preserved on land (see Eyles and Mohajer, 2003; Mohajer et al., 1992). This underscores the value of lake-based seismic risk studies.

This study also furthers understanding of the effects of geologically-recent mid-plate seismicity by specifically identifying the sub bottom geophysical characteristics of lacustrine deformation structures that are the result of non-seismic processes (e.g. ‘kettles; Fig. 13). It highlights, for example, the importance of mass flow during deglaciation. Widespread deformation in lateglacial lacustrine sediments is the product of deposition in highly dynamic proglacial lakes where rapid sedimentation, abrupt changes in lake depth and inflowing meltwater drainage, depositional oversteepening of fan delta slopes and melting of trapped and buried ice all result in deformation. In particular, the study shows that debris flow was a very common process in lateglacial waterbodies in contrast to previous models that emphasize a relatively simple annual control on sedimentation in front of retreating ice margins (Ashley, 1975). Only in the case of the seismically active Lake Timiskaming basin can an earthquake-related origin be suggested on the basis of the unusual dimensions of debris flows in the lateglacial deposits of glacial Lake Barlow-Ojibway (Fig. 15).

The present paper supports the source model for intracratonic earthquakes presented by Bartholomew and Van Arsdale (2012). They suggest reactivation of faulted brittle upper crust has been localized to terrane boundaries and other poorly understood Precambrian basement fractures and lineaments (see also Boyce et al., 2002; Boyce and Morris, 2002; Daneshfar and Benn, 2002; Wallach et al., 1998). It follows that all Precambrian structures are potentially seismogenic. In this regard, the seminal study of Sanford et al. (1985) showed that significant faulting of Precambrian basement structures occurred in southern Ontario during Paleozoic orogenies (Fig. 3, 4); widespread congruence of topographic features such as valleys and lake basins cut into Paleozoic strata, that together with deeply buried Precambrian structures indicate that basement reactivation and attendant fracturing and faulting of overlying strata has been a persistent feature of this part of mid-continent during much of the Phanerozoic (Eyles et al., 1993; Sanford, 1993). Modern drainage patterns cut into glacial sediments are controlled by neotectonic joints (Eyles and Scheidegger, 1999, 1995; Eyles et al., 1997) and stress-release buckles ('pop ups') on the surface bedding planes of Paleozoic strata are common (Jacobi et al., 2007). These various data sets point to an active neotectonic system in this part of mid-continent with the potential for reactivation of deeply buried Precambrian structures close to major population centres. Future work will concentrate on lakes along the strikes of specific Precambrian structures such as the Central Metasedimentary Belt Boundary Zone which underlies Pickering Nuclear Generating Station.

5.7 Conclusions

High resolution seismic reflection profile data sets collected along more than 2000 km of track lines from 16 lakes across Ontario and western Quebec (Gull, Muskoka, Joseph, Rousseau, Ontario, Wanapitei, Fairbanks, Vermilion, Nipissing, Huron, Georgian Bay, Mazinaw, Simcoe, Timiskaming, Kipawa, Parry Sound, Lake of Bays); one of the largest such datasets collected to date in Ontario. A wealth of acoustic data confirms that unconsolidated fine-grained sediments deposited in lakes during and after deglaciation about 10,000 ybp are 'natural seismographs.' Lake floor sediments record ancient earthquakes in the form of co-seismic faults, slumps, debris flows and diapiric water escape structures (collectively referred to as 'seismites') or, alternatively, record the absence of seismic activity where such structures are not present (considered as 'negative' evidence; Strasser et al., 2013).

High resolution sub-bottom seismic records from Parry Sound, Lake Muskoka and Lake Joseph allow identification of lateglacial faulting and slumping likely related to ground shaking during earthquakes triggered by rapid crustal rebound immediately after deglaciation. These locations lie directly above Precambrian terrane boundaries. Lateglacial structures can be readily distinguished from those that affect younger Holocene sediments in lakes Timiskaming and Kipawa along the seismically-active Timiskaming Graben. This area forms part of the Western Quebec Seismic Zone that is associated with the St. Lawrence Rift System, a large failed rift which records ongoing persistent deformation within the North American plate. Deformation is ongoing along the floor of these lakes as indicated by faults, open crevices and recent slumps. Other postglacial deformation structures such as slumps, faults and dewatering structures, which are most readily explained by reference to ground shaking during postglacial earthquakes, occur on the floor of Lake Simcoe, again directly above a Precambrian terrane boundary. This paper supports the model that ongoing mid-plate earthquake activity is a consequence of brittle deformation of the fractured and faulted upper crust of the North American plate. These structures are inherited from phases of obduction and rifting associated with the formation and breakup of the supercontinents Rodinia and Pangea. Compilation of regional earthquake epicentres in eastern North America confirms a distinct broad-scale association with major lineaments, suture zones and failed rifts recording the life histories of Rodinia and Pangea. Acoustic data presented in this paper from sixteen lakes indicate the importance of reactivated Precambrian terrane boundaries that are exposed at surface on the Shield and underlie many lake basins but which southward become deeply buried under thick covers of Paleozoic and Pleistocene glacial sediments beneath the principal urban centres of eastern Canada. The precautionary principle suggests that all Precambrian structures in mid-continent North America must be considered potentially seismogenic. The enormous potential impact of mid-continent earthquakes on the economy of this heavily populated area (e.g. AIR Worldwide, 2013) makes further study a high priority and underscores the need for additional regional-scale investigations of natural lacustrine seismographs.

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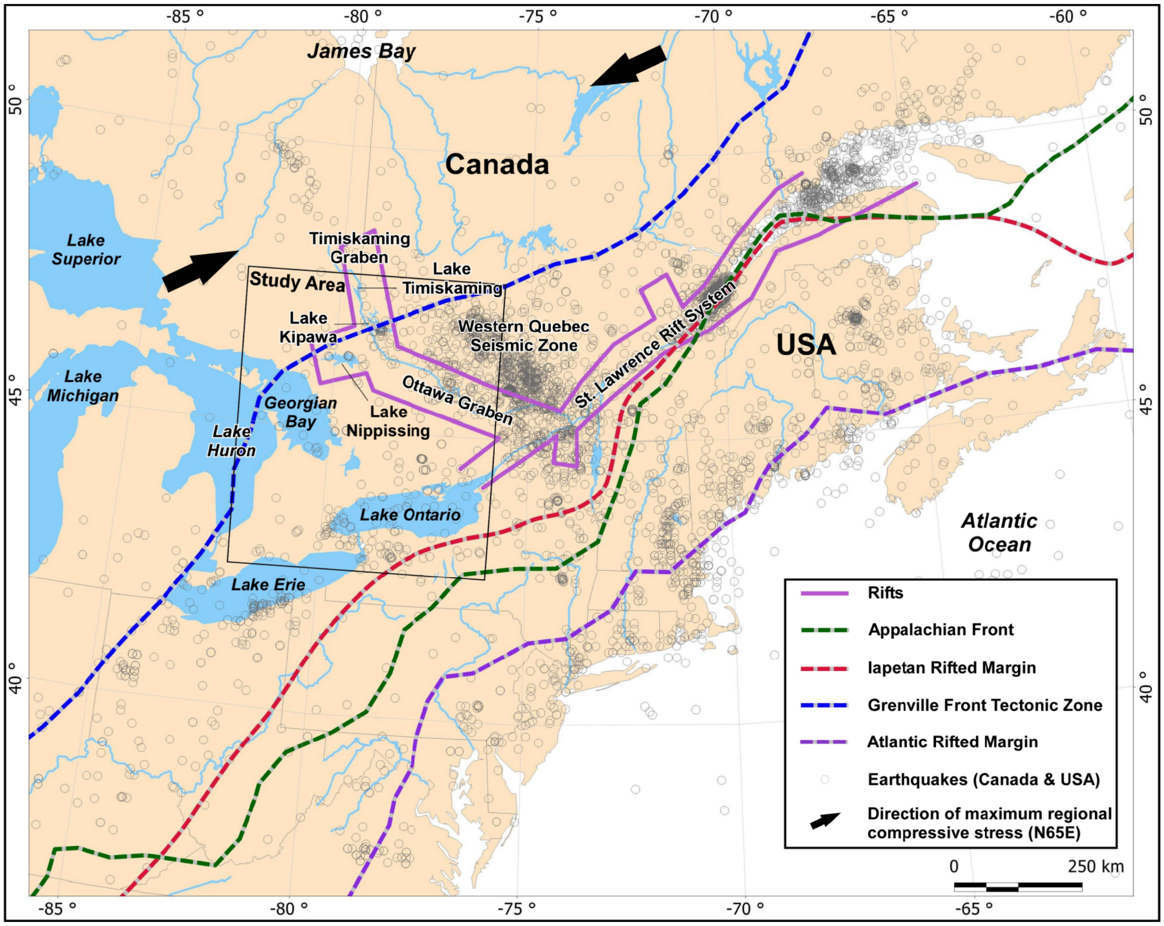


Figure 1: Broad scale structure of eastern North America resulting from formation and breakup of Rodinia and Pangea with historic earthquake epicentres (1627 to 2013 AD); after Natural Resources Canada (2013) and Halchuk (2009). Major structural boundary locations were adapted from Bartholomew and Van Arsdale (2012), Boyce and Morris (2002), Mazzotti (2007), Thomas (2006), Withjack et al. (2002) and Carr et al. (2000). Direction of maximum regional compressive stress (for the study area) from Nuclear Waste Management Organization and AECOM Canada Ltd. (2011).

Figure 2: (Next Page) Terrane boundaries within the Grenville Province showing lakes where (potential) seismites have been identified in lateglacial or postglacial deposits (red stars). Lakes Nipissing, Timiskaming and Kipawa lie along prominent failed rift structures associated with the St.Lawrence Rift System (Fig. 1). Structural data summarized from Bartholomew and Van Arsdale (2012), Mazzotti (2007), Boyce and Morris (2002) and Carr et al (2000). Letter designations after Carr et al (2000).

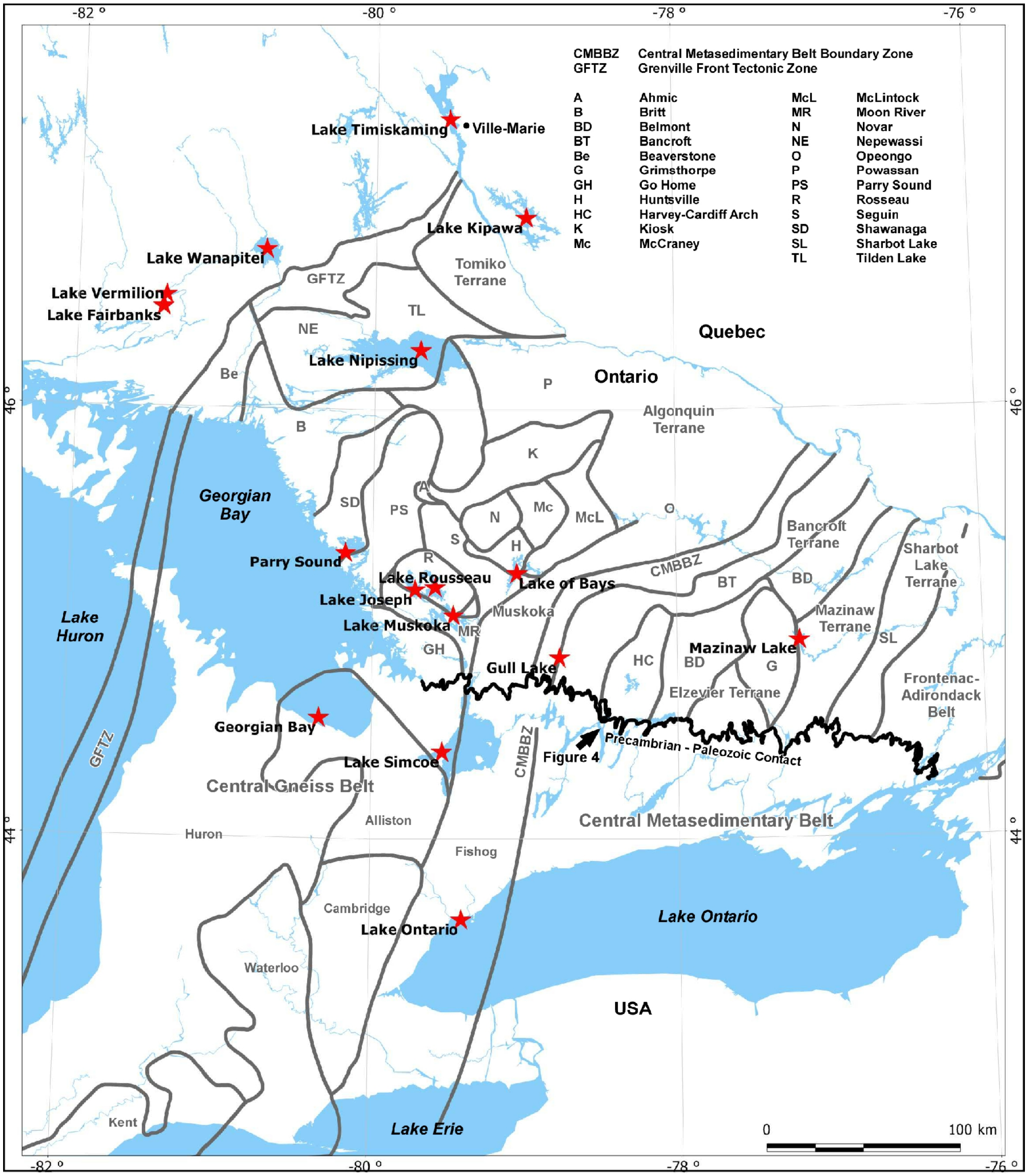


Figure 2: Continued ...

Figure 3A: (Next Page) Study area. Studied lake basins (red stars) are indicated along with earthquake epicentres (Natural Resources Canada, 2013; Halchuk, 2009), faults, lineaments and terrane boundaries (Ontario Geological Survey, 2011; Armstrong and Carter, 2010; Boyce and Morris, 2002; Daneshfar and Benn, 2002; Atomic Energy Control Board, 2000; Carr et al., 2000; McFall and Allum, 1989; Sanford, 1993).

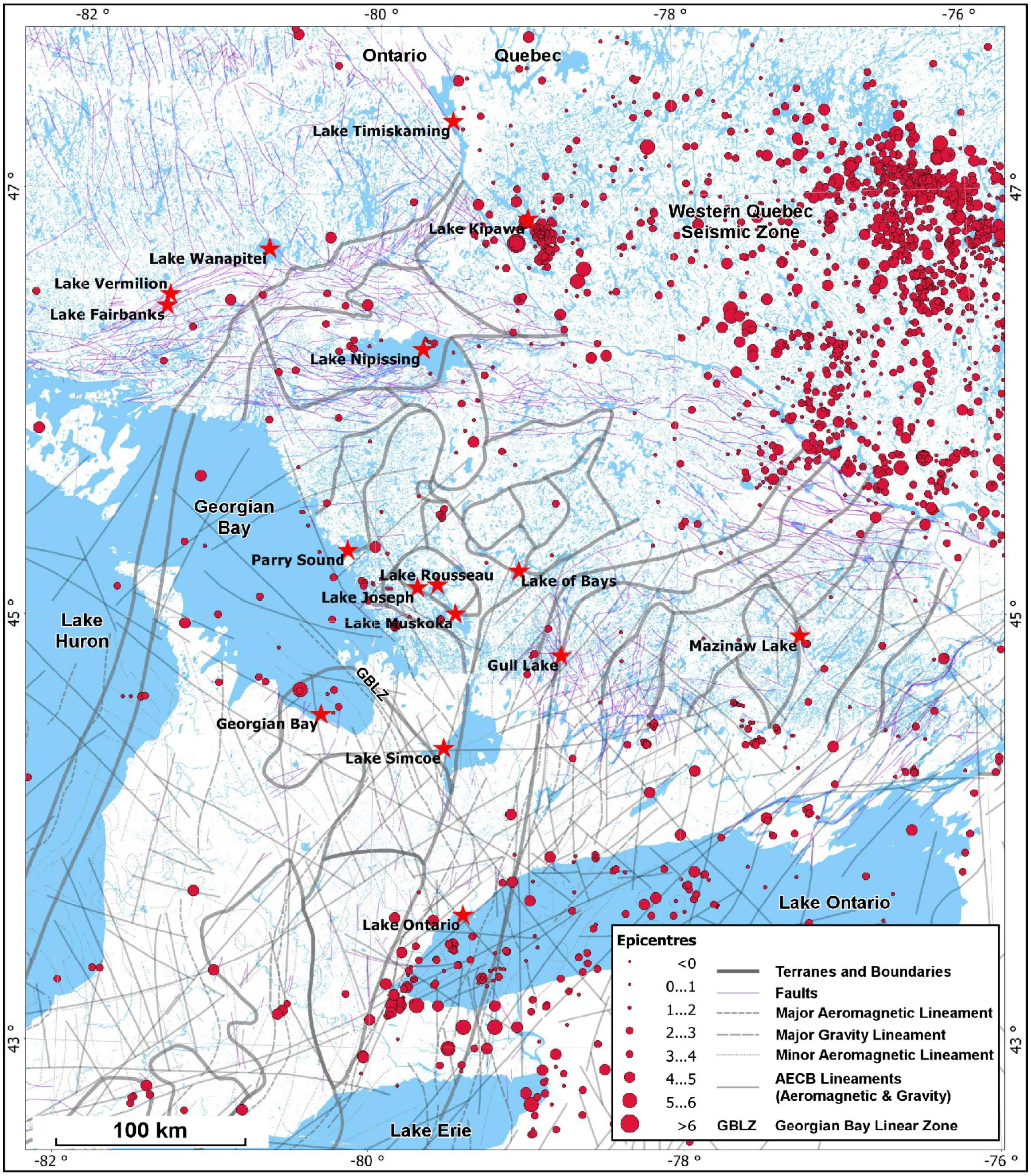
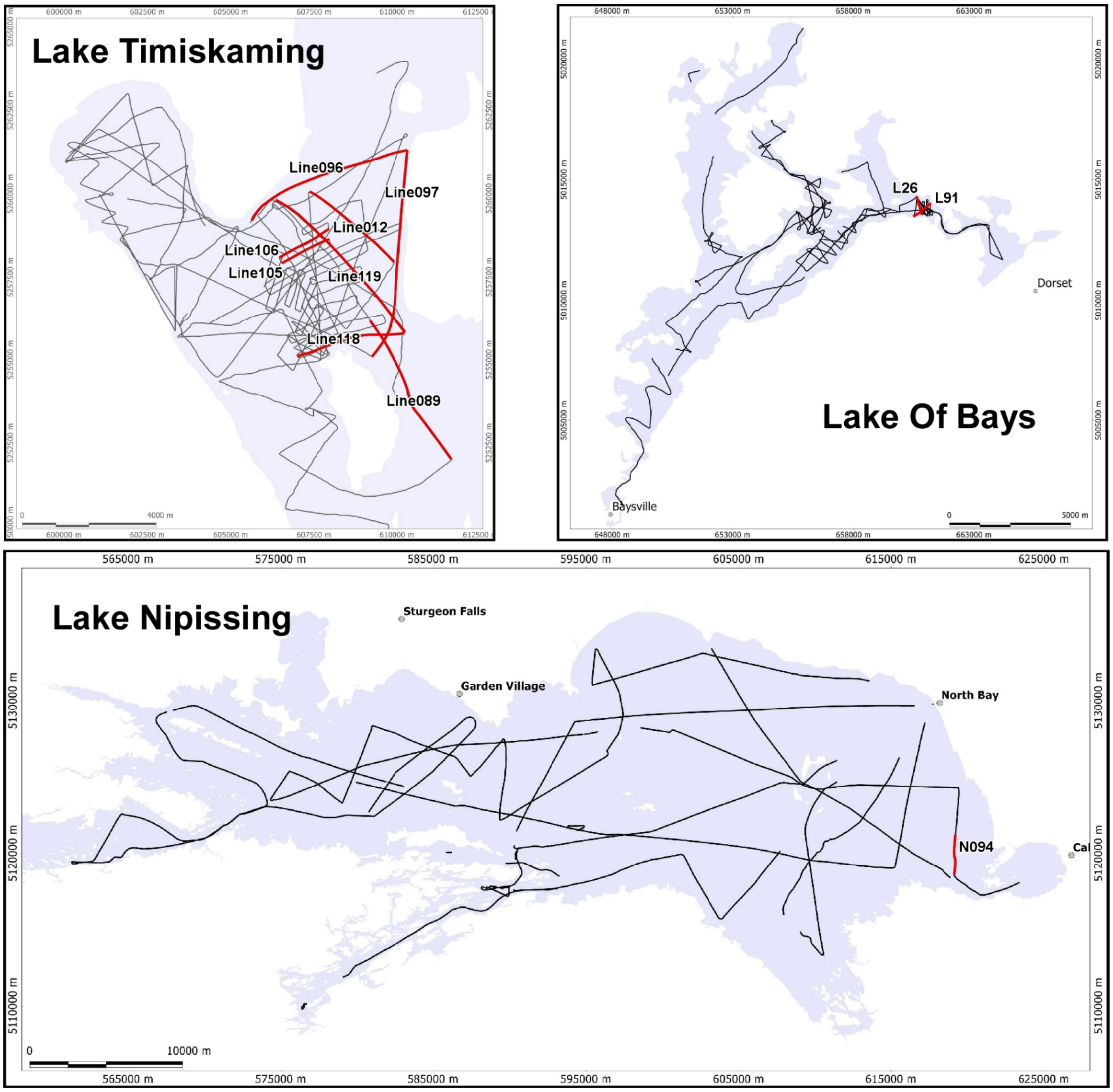
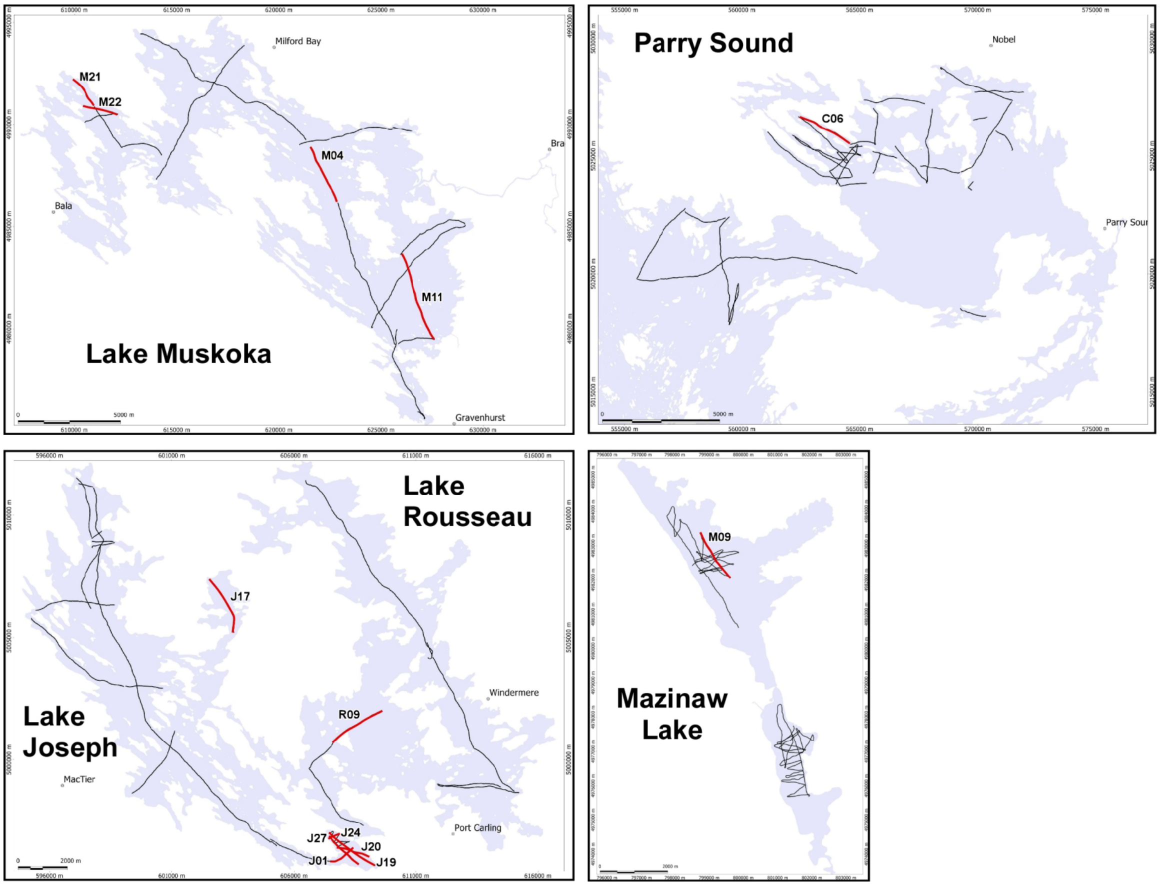


Figure 3A: Continued ...



Figures 3B and C: Seismic data collection track lines for lakes shown in Figure 3A; lines used in this paper are highlighted in red and numbered.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M22
M04
M11
Bala
Gravenhurst
Parry Sound
Nobel
C06
Lake Rousseau
Lake Joseph
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R09
Windermere
MacTier
Port Carling
J27
J24
J20
J01
J19
Mazinaw Lake
M09

Figure 3C

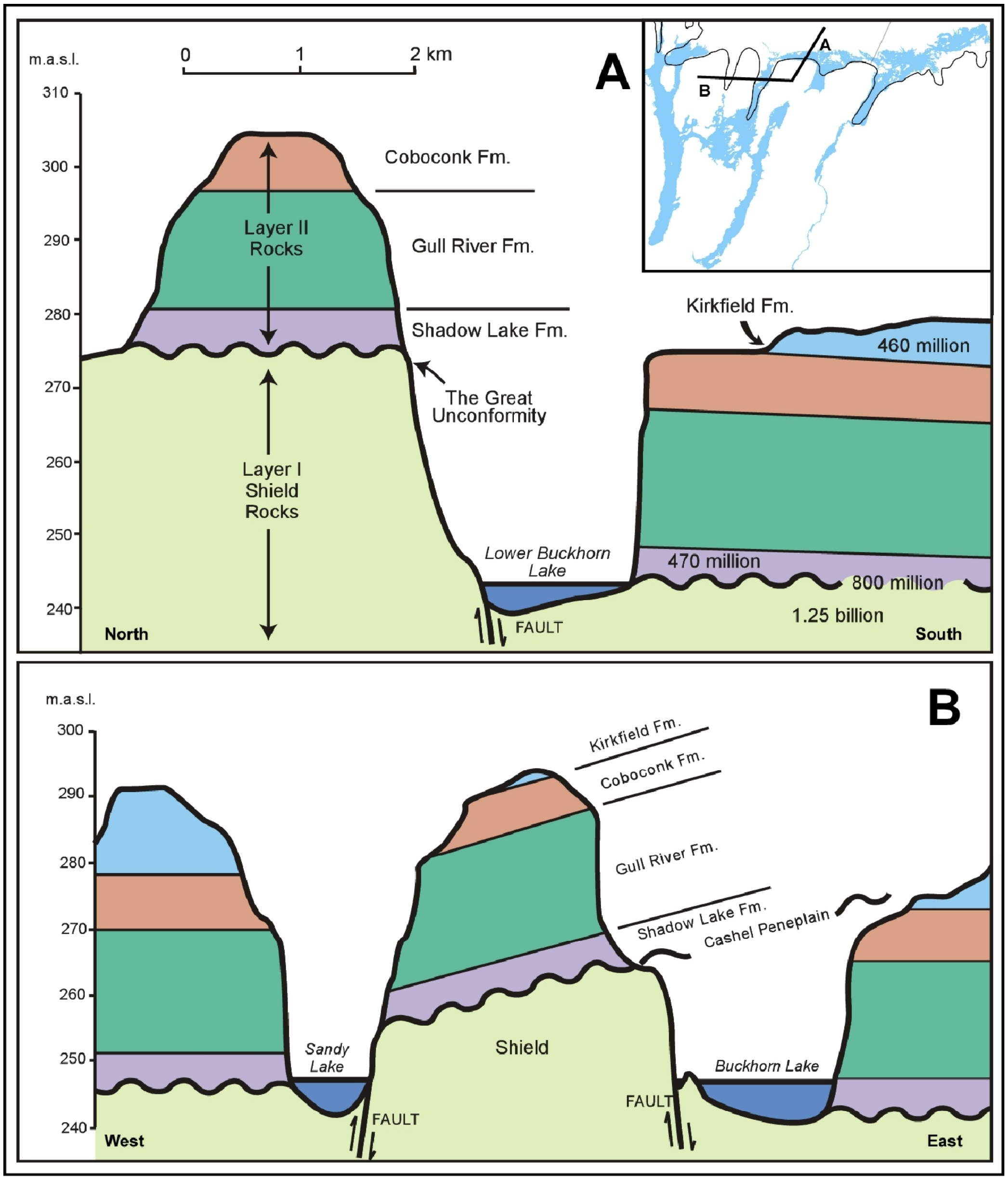

Figure 4: Schematic geological sections in the vicinity of Buckhorn Lake in south-central Ontario (modified from Sanford, 1993). Faulting of Paleozoic strata indicates reactivation of Precambrian structures. Location of sections is indicated on inset map and on Figure 2 (along with the location of the Precambrian-Paleozoic contact, in bold).

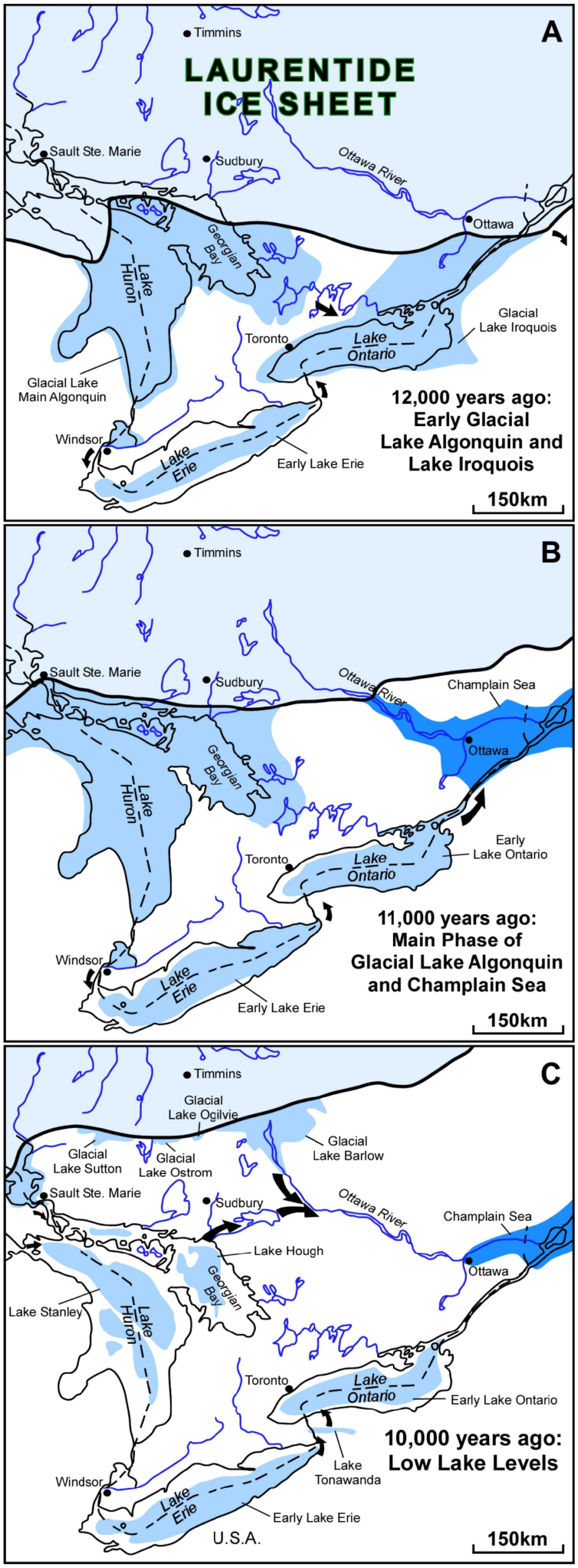

Figure 5: Phases of lateglacial and postglacial lakes and marine incursions in Ontario (after Jackson et al., 2000). Arrows indicate lake outflows.

Figure 6: (Next Page) Interpretation of seismic line 097 from Lake Timiskaming (see Figure 3B for location).. Prominent debris flow deposits (debrites) occur within lateglacial Lake Barlow-Ojibway deposits. Panels are organized A-B-C, left-to-right (North-to-South). Panel B2 is an inset (close-up) from Panel B1. Shaded areas represent interpreted debris flows during the postglacial (PG – pink) and lateglacial (LG – blue).

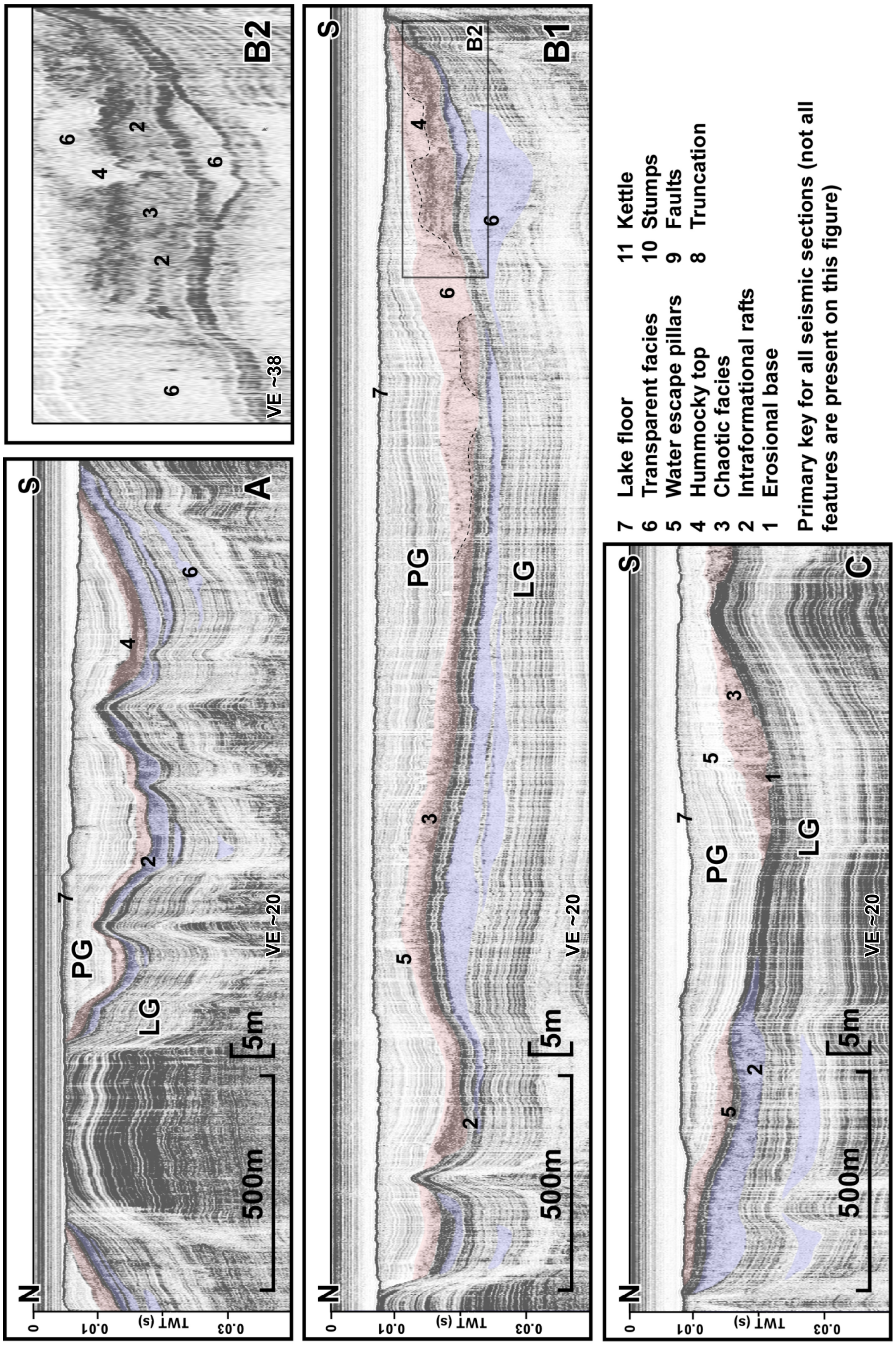

Figure 6: Continued ...

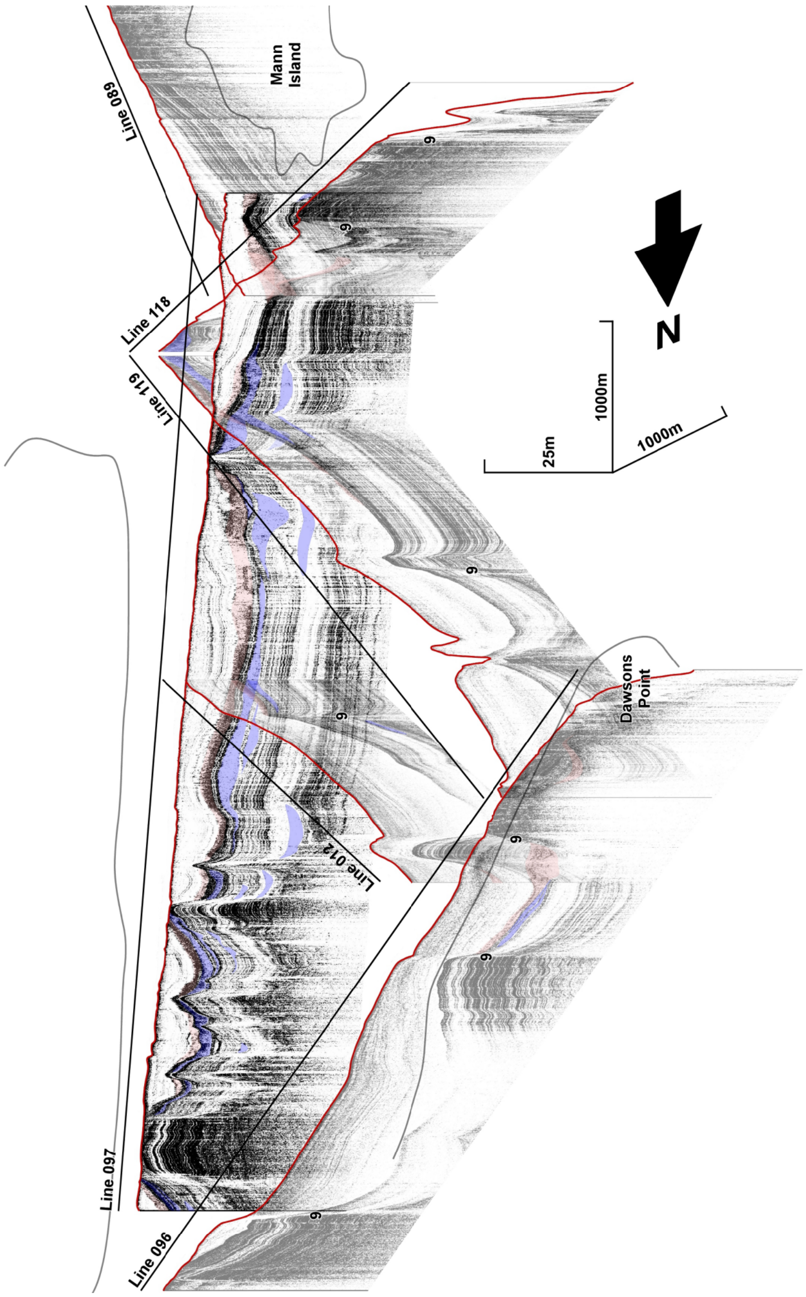

Figure 7: Fence diagram of seismic facies in Lake Timiskaming showing broad extent of prominent debris flow facies seen as massive transparent beds (‘c-m-s facies’; see text for details) within lateglacial Lake Barlow-Ojibway deposits represented here as shaded units (refer to Fig. 6 for details). Refer to Fig. 3B for location for seismic lines.

Figure 8: (Next Page) Debris flow (features 1, 3 and 6) and fluid escape structures (5) in Lake Rosseau (A); debris flows (1-4 and 6) in Lake Joseph and Lake Muskoka (B, C); overpressuring of debris flow facies (6) and subsequent water escape structures (5) is commonly seen in many lakes such as in Lake of Bays (D). Refer to Fig. 6 for numeric code descriptions. Refer to Figs. 3B-C for location of seismic lines.

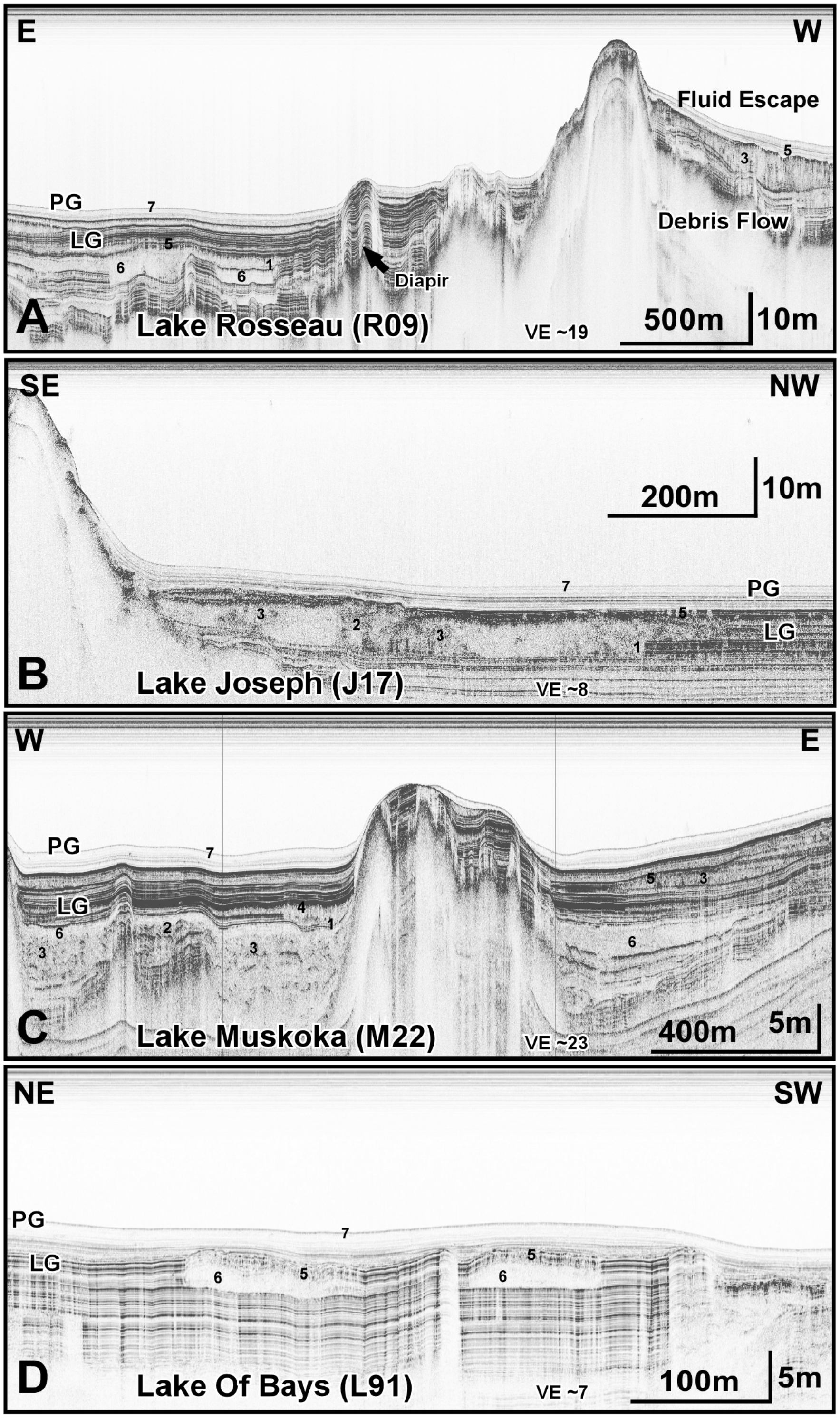

Figure 8: Continued ...

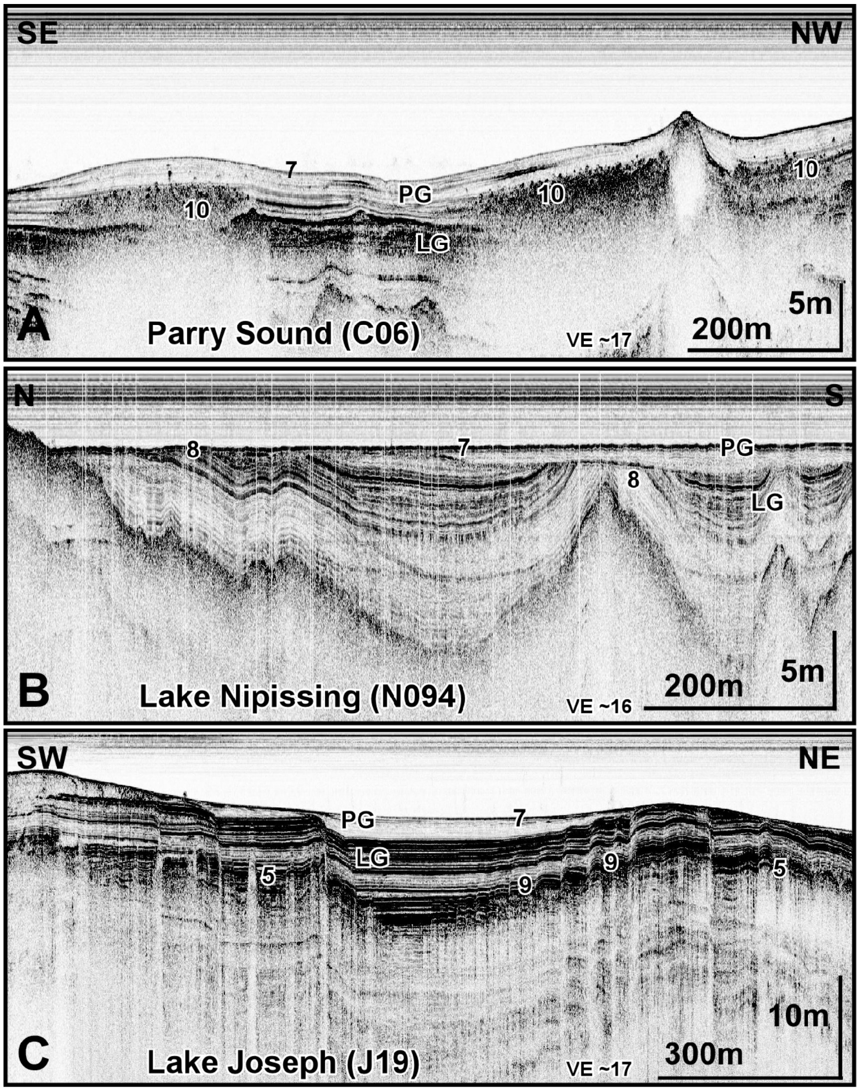

Figure 9: (A) Selected seismic profiles from: Parry Sound on Georgian Bay showing contact between lateglacial and postglacial sediment with horizons interpreted as stumps of trees (10) that grew at times of lowered lake level and were subsequently killed by rising water levels after 8000 ybp. (B) Prominent erosional surface truncating lateglacial sediment in Lake Nipissing (7/8). (C) Faulted lateglacial sediment in Lake Joseph (9) with fluid-escape structures (5). Refer to Figs. 3B, C for location of seismic lines and Figure 6 for numeric code descriptions.

Figure 10: Outcrop of folded and faulted glacial Lake Barlow-Ojibway silty-clays near Ville-Marie, Quebec (east of Lake Timiskaming; refer to Figure 2 for location). These sediments were probably deformed by co-seismic shaking. See text for details

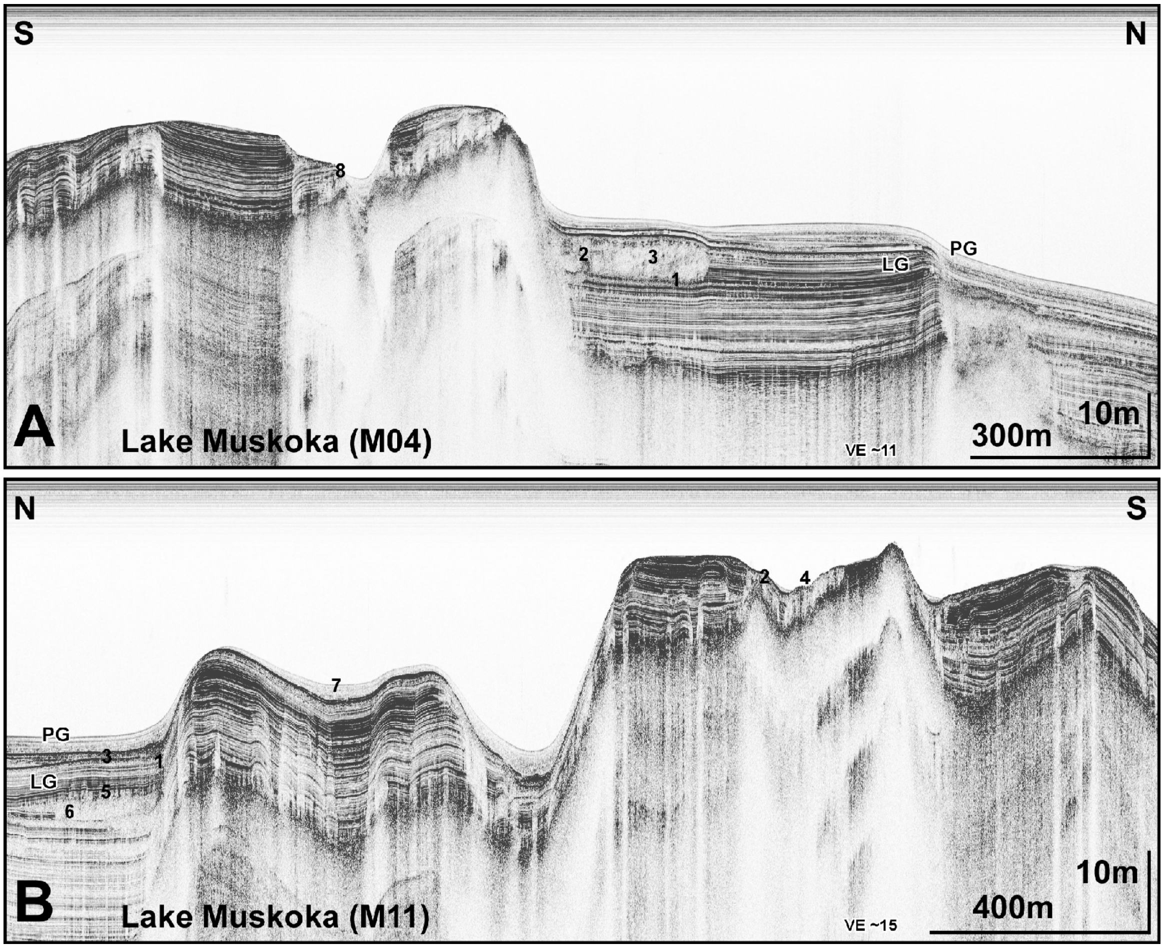

Figure 11: Seismic lines from Lake Muskoka showing debris flow features (1-4 and 6). Refer to Fig. 6 for numeric code descriptions. Refer to Figs. 3C for location of seismic lines.

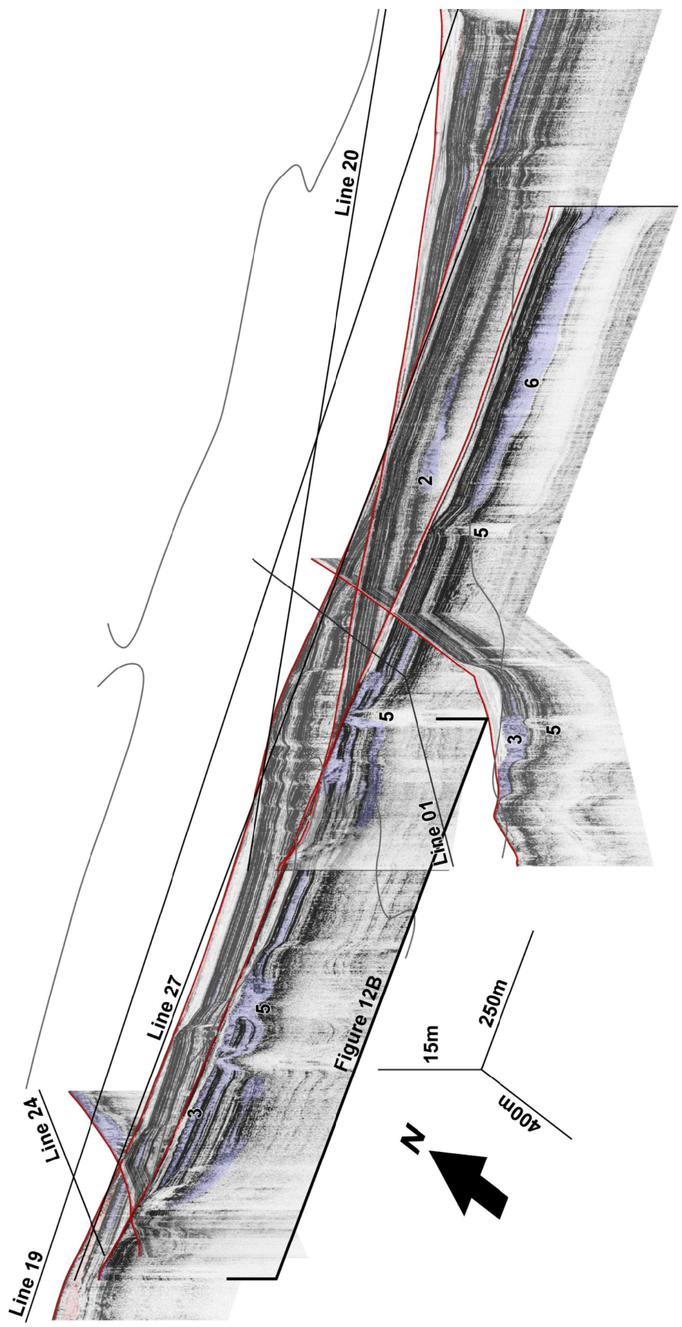

Figure 12A: Fence diagram view of seismic facies in Lake Joseph showing extent of debris flow facies (shaded and 2, 3 – refer to Fig. 6 for numeric codes and colour descriptions) and vertical pipe-like structures (5; see also Fig. 12B for a detailed section, location indicated). Refer to Figs. 3B for location of seismic lines.

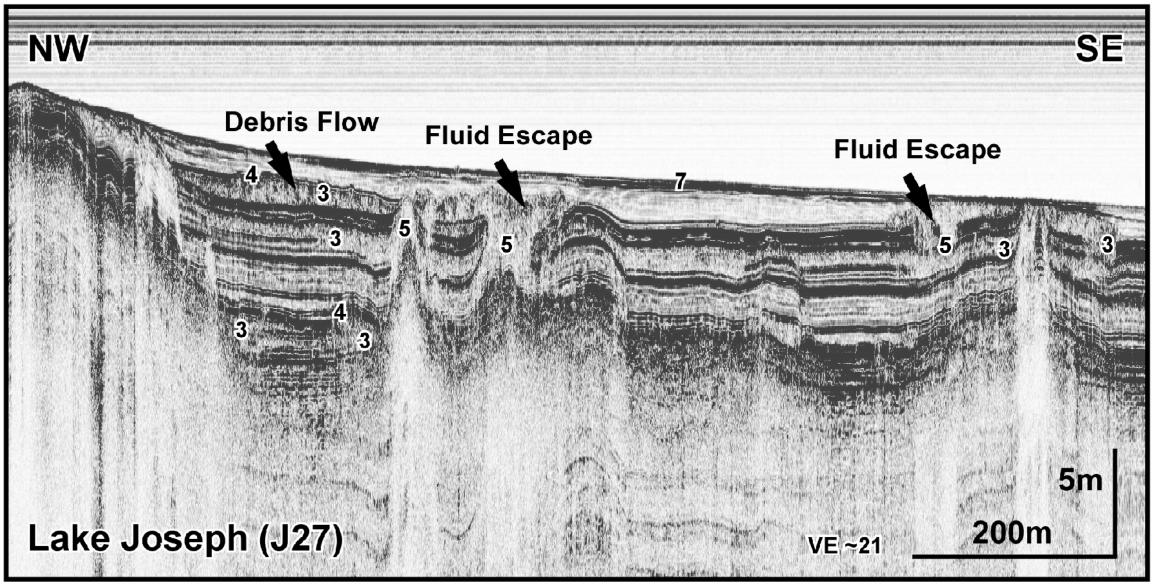

Figure 12B: Seismic line (J27) showing fluid escape structures in lateglacial fill of Lake Joseph recorded by vertical pipe-like structures (5) where overpressured sediment has moved upward. Note correlative hummocky debris flow bed (4) at left suggesting mass flow and water escape were coeval events likely triggered by ground shaking. The lake lies directly above a Precambrian terrane boundary (Fig. 2). Refer to Fig. 6 for numeric code descriptions. Refer to Fig. 3B for location of seismic line.

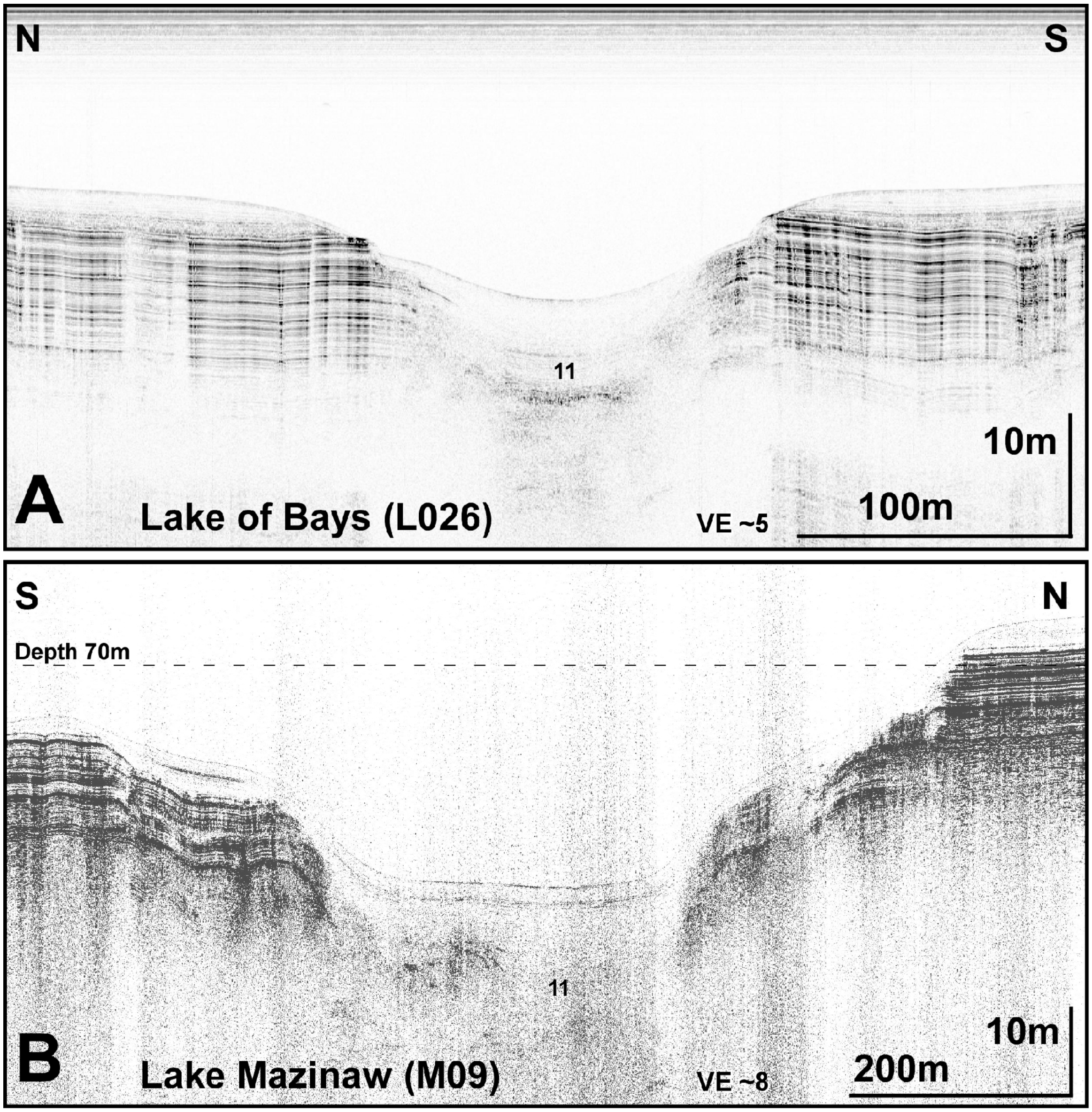

Figure 13: Seismic lines illustrating craters (kettle holes, 11) on the floor of Lake of Bays (A) and Lake Mazinaw (B) and) resulting from melt of large ice blocks buried by sediment during deglaciation. Refer to Fig. 6 for numeric code descriptions. Refer to Fig. 3B for location of seismic lines.

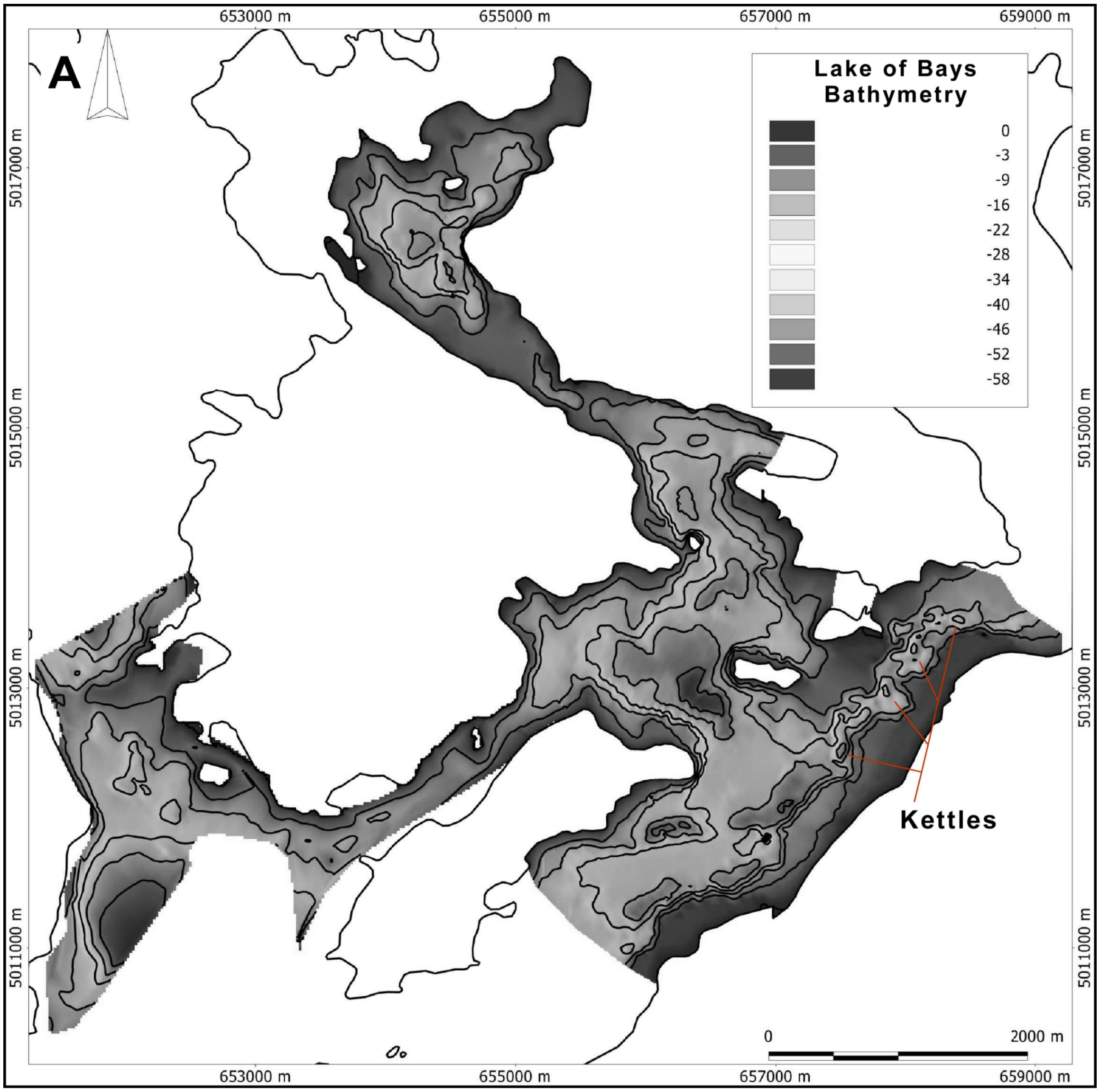

Figure 14. Bathymetry of Lake of Bays (A) and Lake Mazinaw (B; adapted from Eyles et al, 2003) showing prominent craters originating as kettle holes (Fig. 13). Bathymetric data collected using a dual-channel bottom sounder with line spacing of 10-50 m (fine to medium resolution) through to 100-150 m (coarse resolution) with crossing (tie) lines.

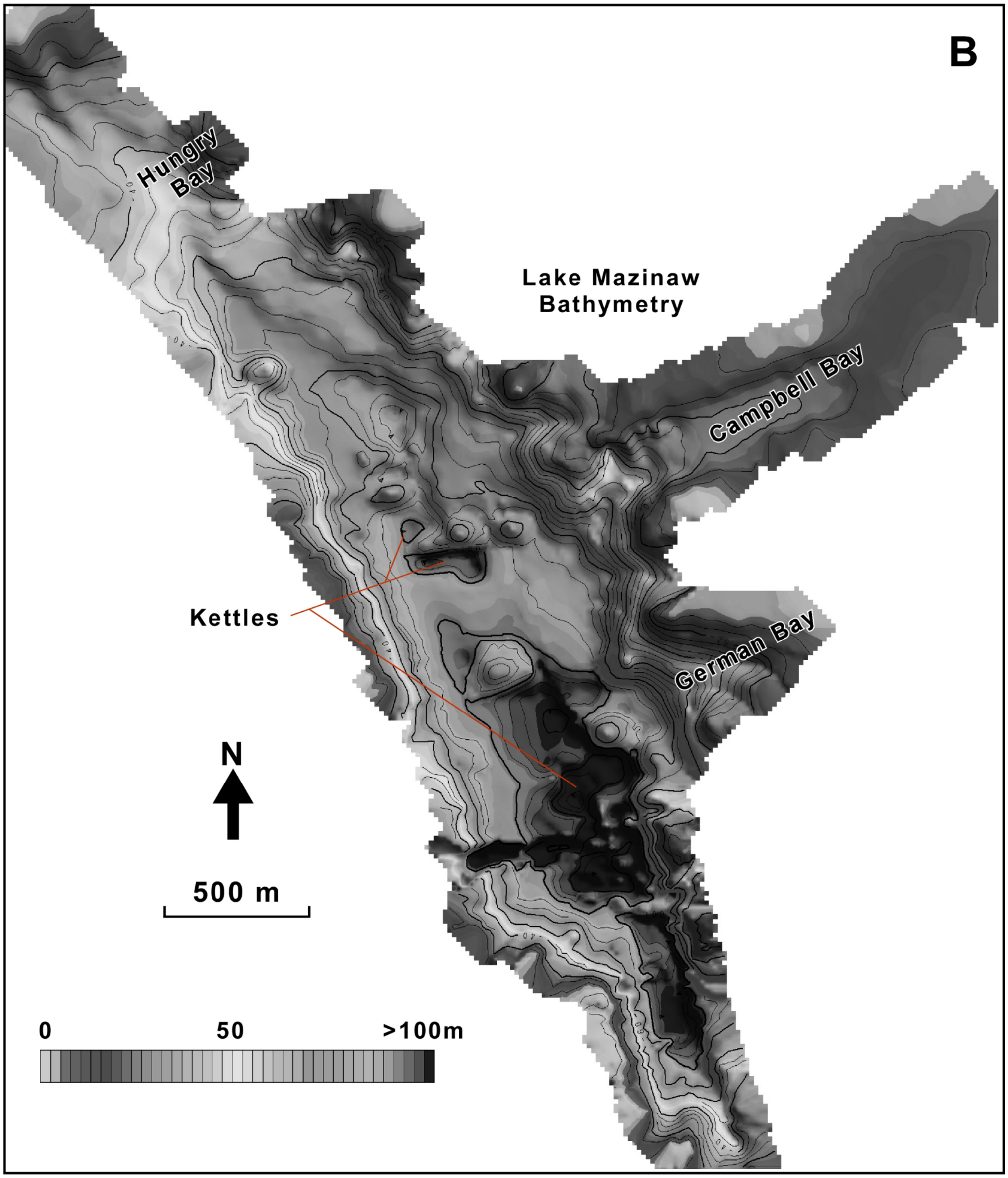
B
Hungry Bay
Lake Mazinaw Bathymetry
Campbell Bay
Kettles
German Bay
N
500 m
0
50
>100m

Figure 14B

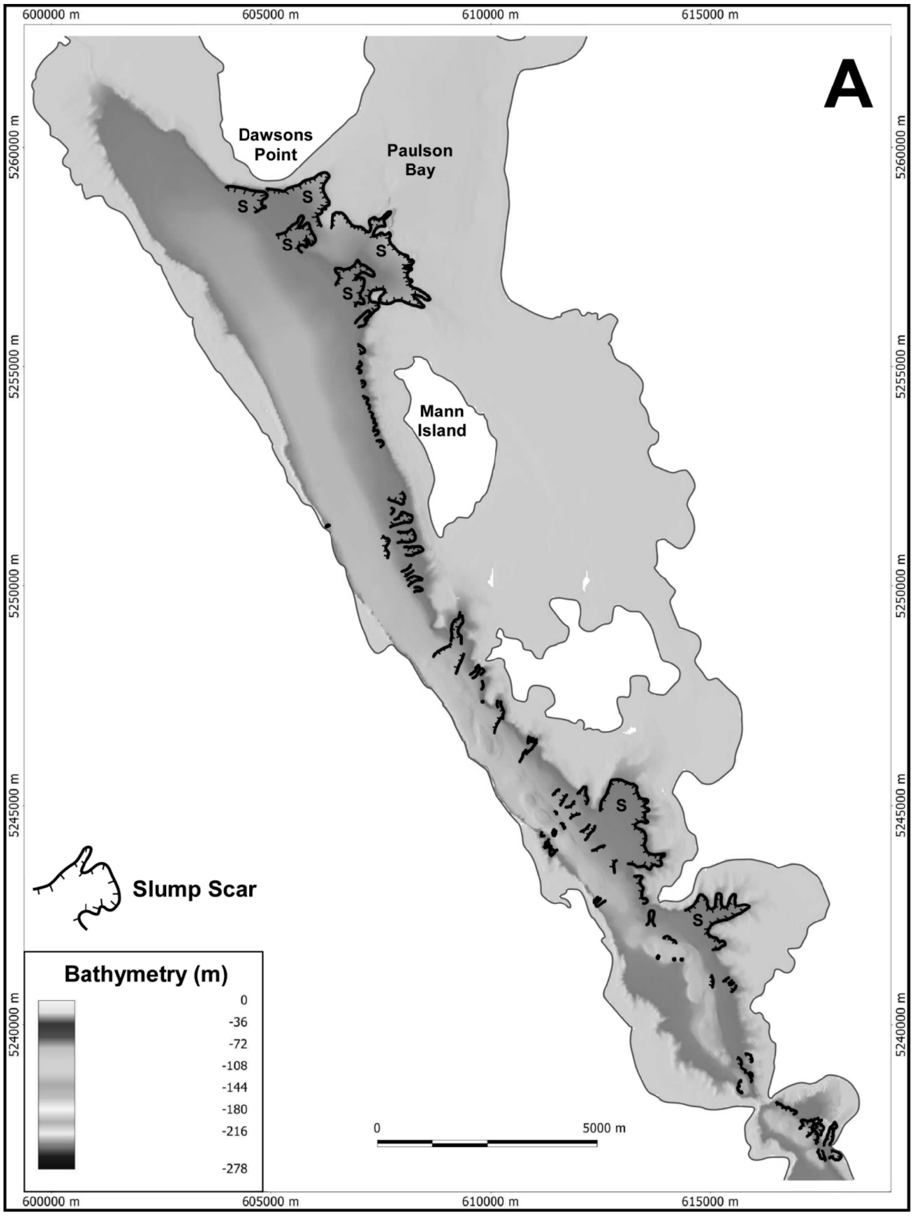

Figure 15 (A-B) Multi-beam bathymetry of Lake Timiskaming showing distinctly crenellated form of slope around basin margin created by bowl-shaped slump scars (marked 'S') produced by spring sapping and mass wasting. Debris flow lobes extend basinward from mouths of slump scars (Fig. 16). Bathymetric data from Canadian Hydrographic Survey (2005).

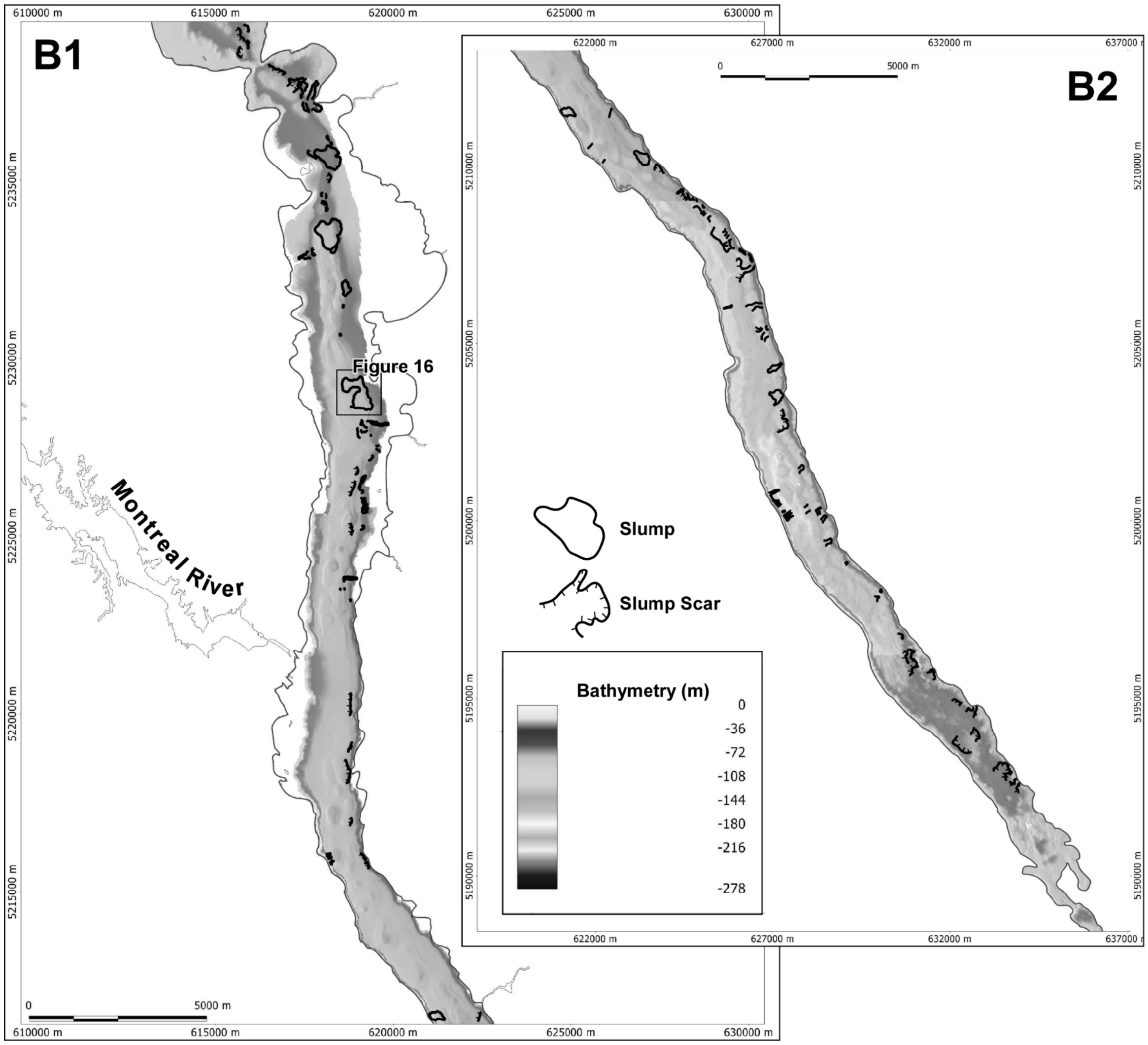
B1
B2
Figure 16
Montreal River
Slump
Slump Scar
Bathymetry (m)
0
-36
-72
-108
-144
-180
-216
-278
0
5000 m
610000 m
615000 m
620000 m
625000 m
630000 m
622000 m
627000 m
632000 m
637000 m

Figure 15B

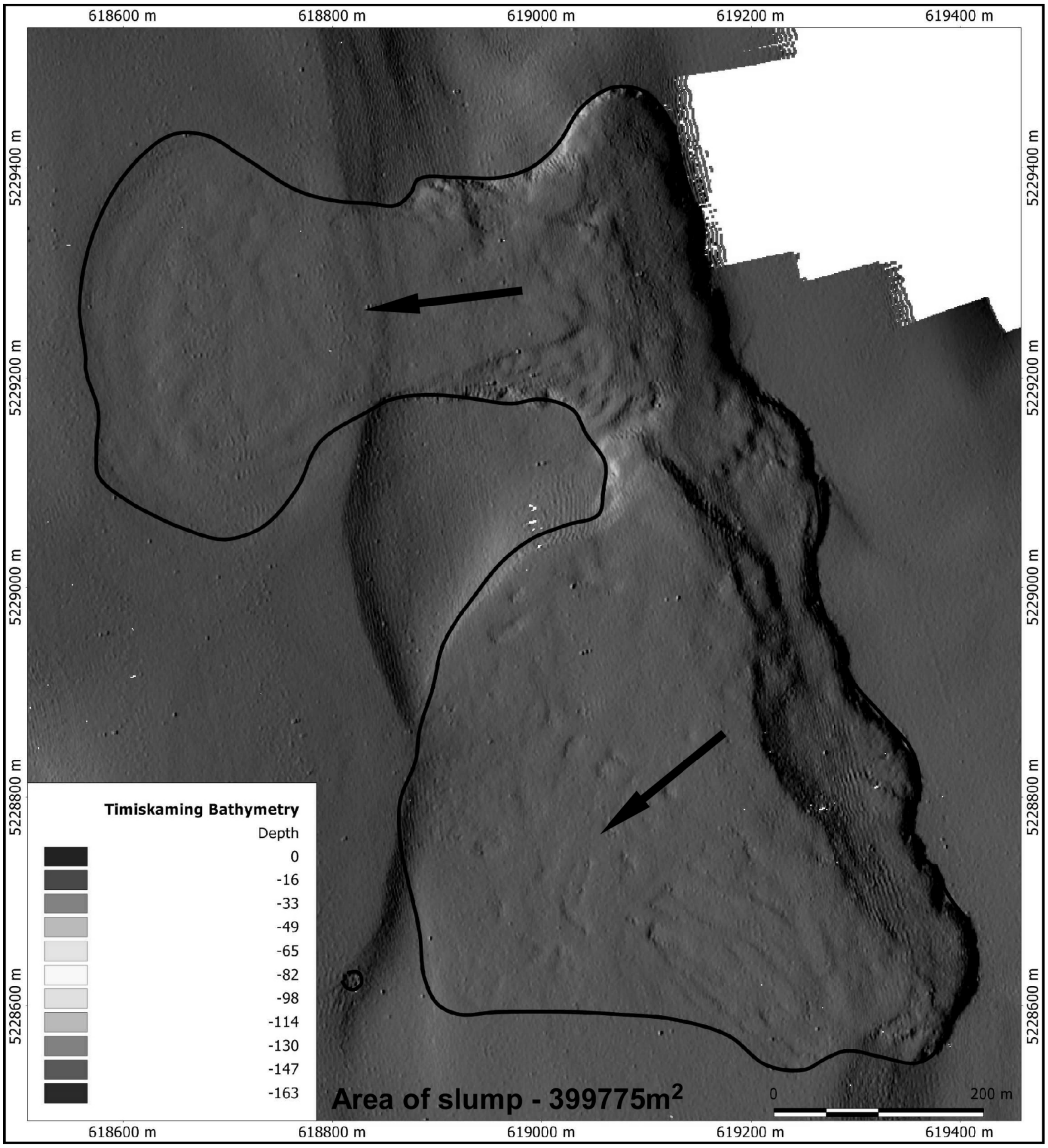

Figure 16: Debris flow lobes (outlined; imaged from multi-beam bathymetry) with hummocky upper surfaces at the foot of the eastern basin margin slope of Lake Timiskaming (see Figure 15 B1 for location). Debris flow activity was a common process during deglaciation of Ontario lakes (Figs. 8, 11) but the largest debris flow lobes identified to date occur in Lake Timiskaming (Fig. 6).

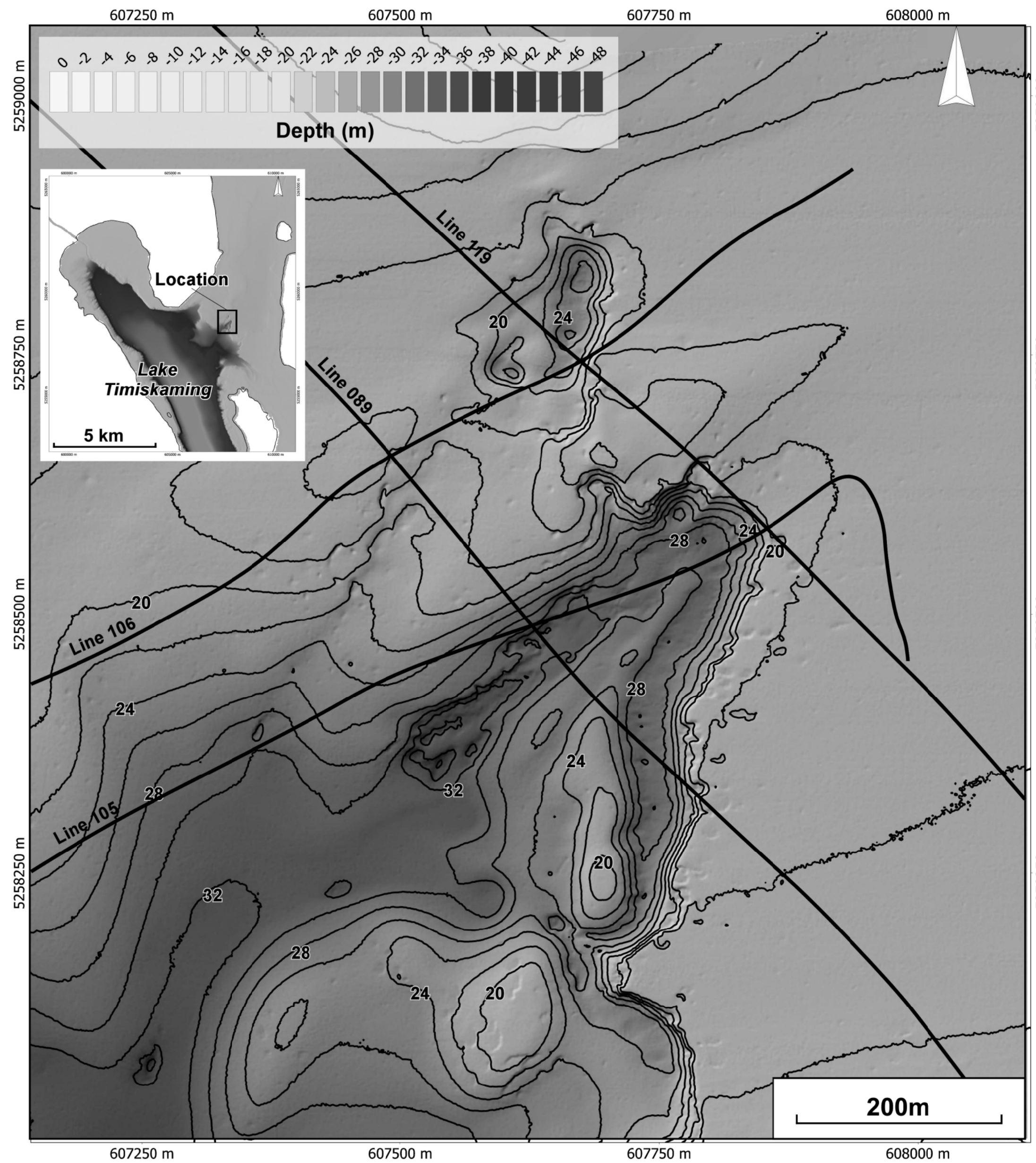

Figure 17: Bathymetric data showing gullies on floor of Paulson Bay in Lake Timiskaming resulting from spring sapping and collapse over faults. Seismic profiles (seismic track lines numbered and indicated by coarse lines) suggest upward movement of overpressured sediment below the area (Fig. 18). Note enclosed collapsed depression to the north directly over underlying fault. Refer to inset for location.

Figure 18: (Next Page) High resolution seismic profiles in Lake Timiskaming across area shown in Figure 17 showing pillars of deformed lateglacial sediment identified as diapirs; see Figure 8 for similar structures in Lake Rosseau. Refer to Figs. 3B and 17 for location of seismic lines and Fig. 6 for numeric code descriptions.

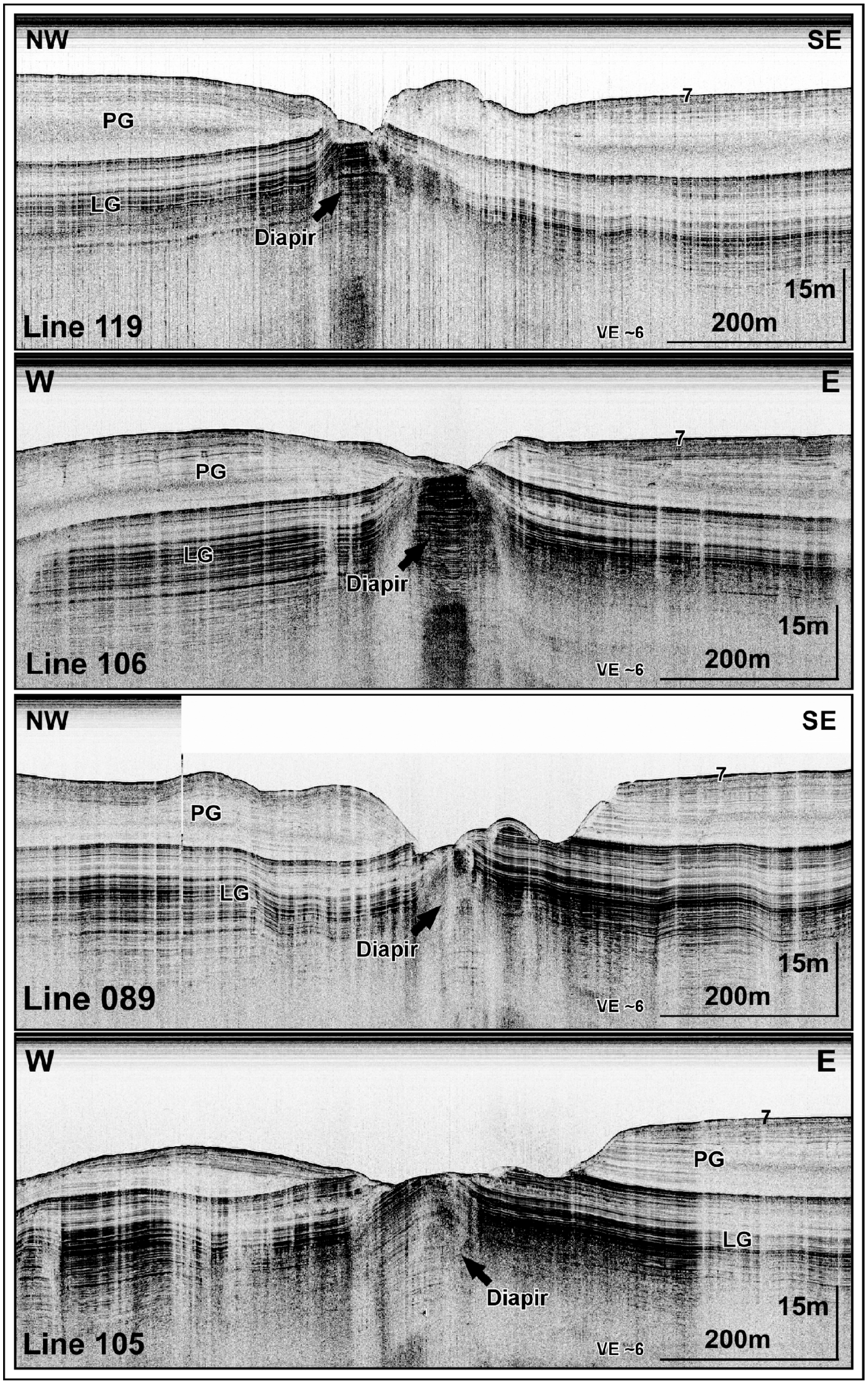

Figure 18: Continued ...

Figure 19: (Next Page) Multi-beam images of open fractures on floor of Lake Timiskaming that lie directly above faults and which suggest ongoing neotectonic activity and structurally controlled subsidence. Location indicated on D. Enclosed depressions (marked 'ED' on C) similar to those depicted here are also seen on Fig. 17. Approximate look direction and position for A1-C1 is indicated (arrowed) on A2-C2. Features in A and B are approximately 1m in height. The enclosed depression in C is approximately 2m in depth with the scarp (to the east) approximately 10m high. Water depths vary between 15 and 30m.

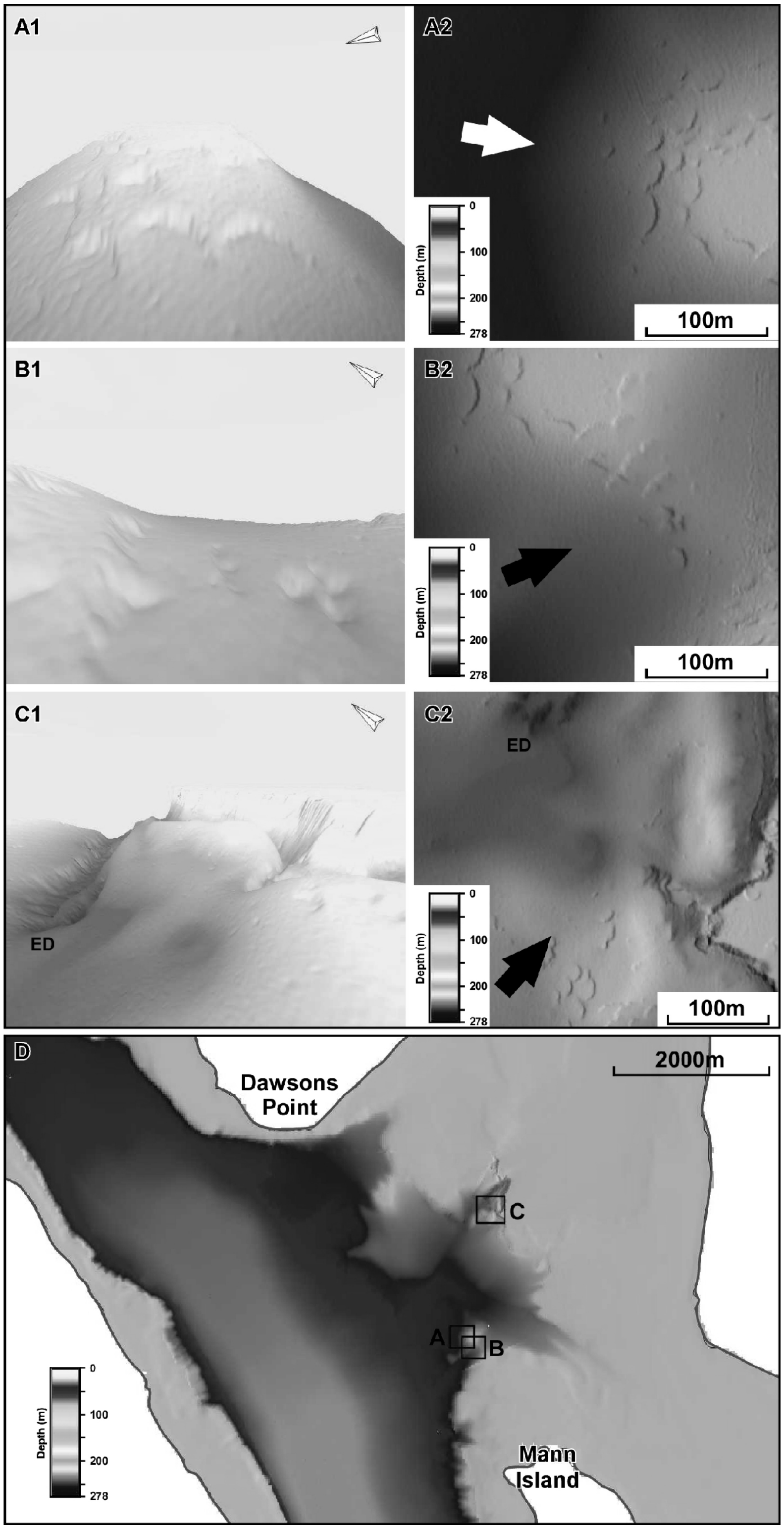

Figure 19: Continued ...

Figure 20: (Next Page) Total field magnetic map of Kempenfelt Bay in Lake Simcoe (Fig. 2) .Northwest-trending linear anomaly marks boundary of Alliston-Go Home terrane in Precambrian basement. Boundary is offset by left-lateral slip of approximately 500 m on southwest-trending fault. (B) Bathymetric map of basin. Note debris flow lobe (arrowed) on southeastern margin of the shear zone. Adapted from Boyce and Pozza (2004) and Boyce et al. (2002).

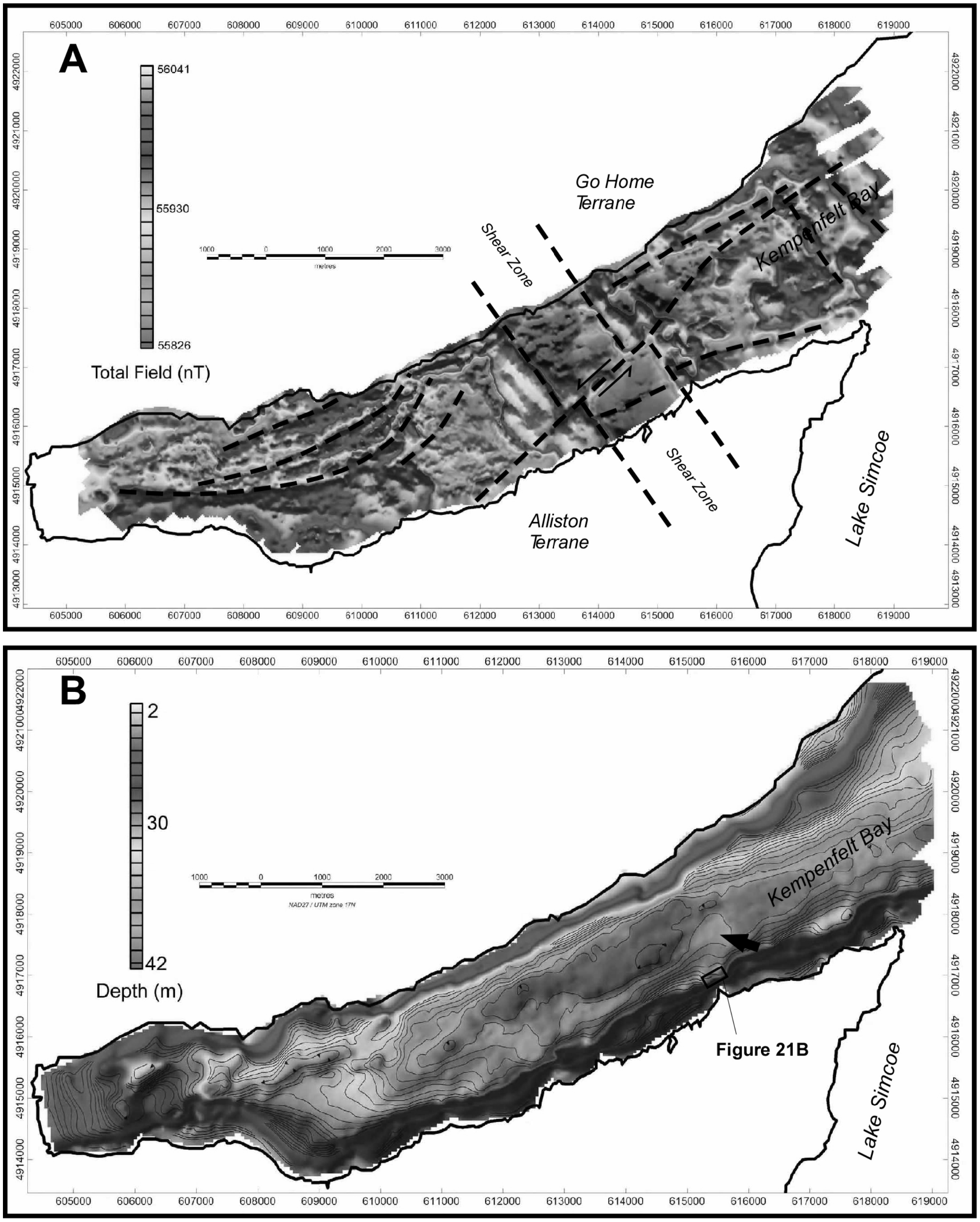

Figure 20: Continued ...

Figure 21: (Next Page) (A) Sub-bottom sonar profile showing normal faulting of lake bottom Holocene muds and silts in Kempenfelt Bay. Graben is filled by undisturbed Holocene sediment. (B) Side-scan sonar image of southern basin margin of Kempenfelt Bay, showing large rotated slump block of Holocene sediment (location of image shown in Fig. 20B) on the southeastern margin of the Alliston-Go Home shear zone. Slumping is interpreted as a co-seismic feature associated with basement fault reactivation. Adapted from Boyce et al. (2002).

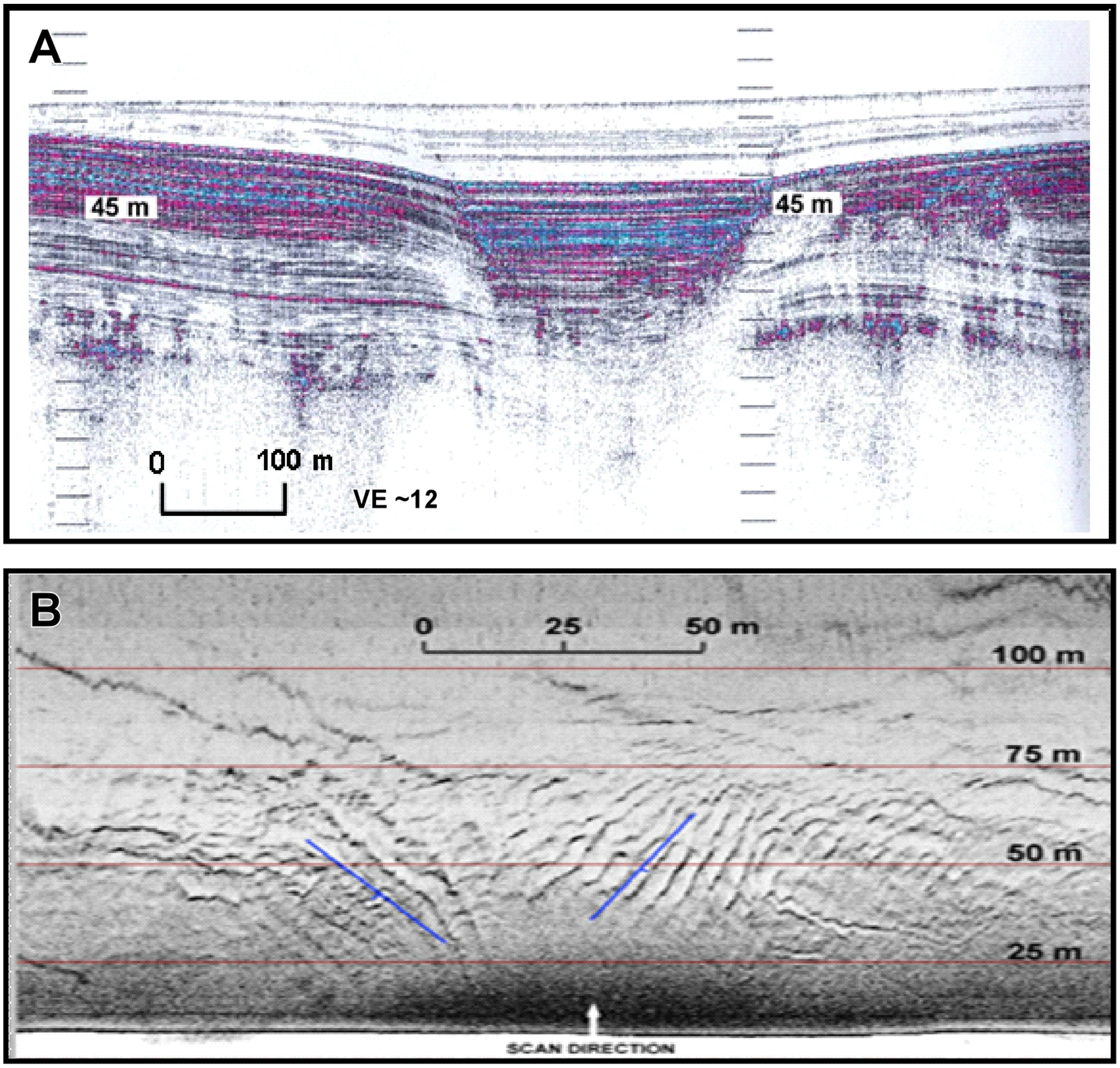

Figure 21: Continued ...

Figure 22: (Next Page) (A) Sub-bottom sonar profile across small sub-basin showing gas escape columns (arrowed). Escape of gas (or pore water) produces small pits and linear trough-like features (inset) on lake floor. Troughs suggest movement of gases along fracture planes. (B) Sub-bottom profile showing 'en echelon' reverse faults in postglacial laminated sediments. Note rotation of fault blocks and small vertical offsets of bedding (1-2 m). Structures indicate compressional deformation of the lake-floor, possibly in association with strike-slip movements on basement faults identified in Figure 20B. Adapted from Boyce et al. (2002).

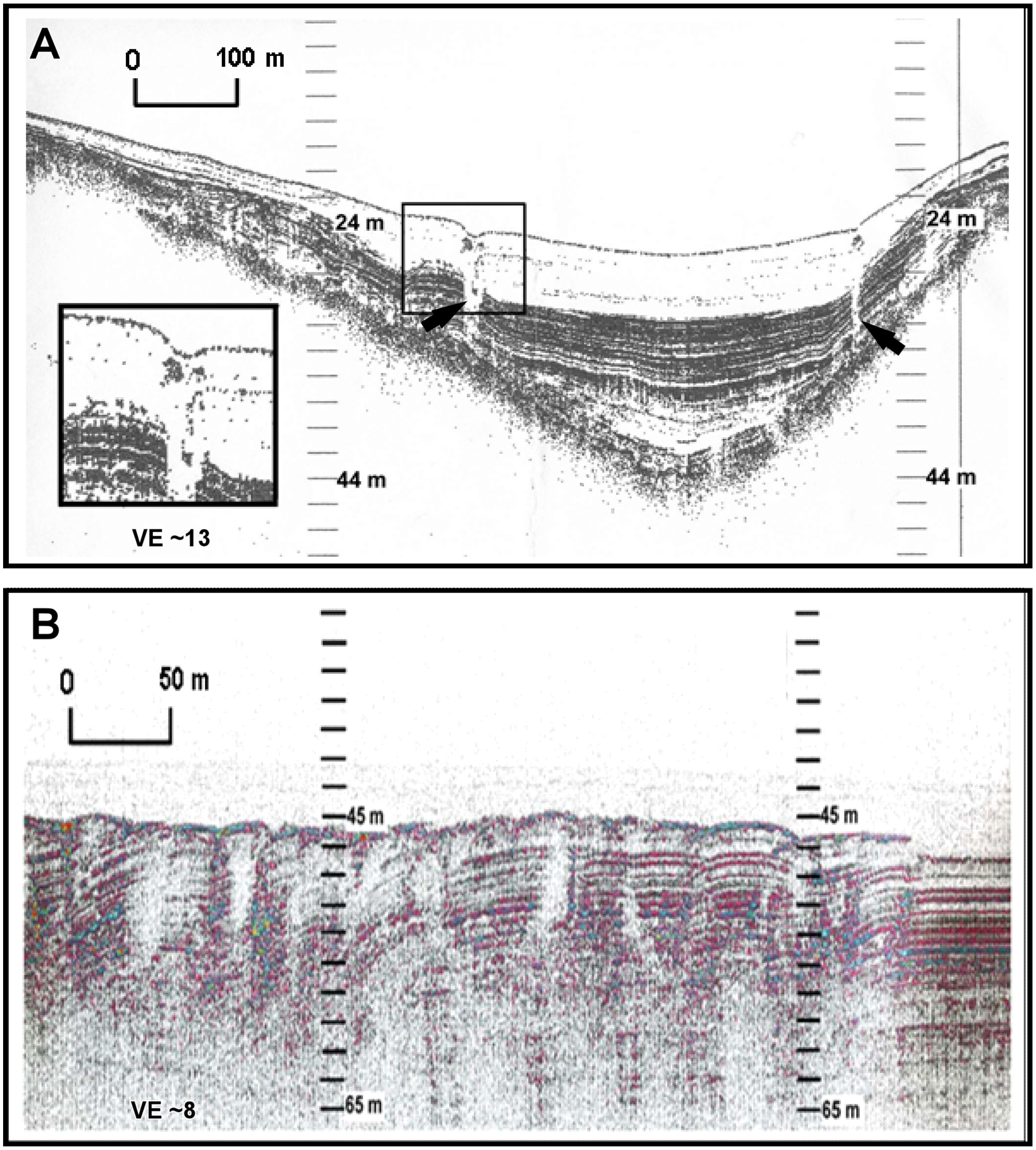

Figure 22: Continued ...

Figure 23: (Next Page) Location of postglacial and lateglacial seismite features identified in this study with major Precambrian tectonic boundaries (grey lines) and earthquake epicentres (pink circles). Structures reported by other authors are also shown: (1) Todd and Lewis, (1993) and Boyce et al (2002); (2-4) Thomas et al. (1993); (3) Jacobi et al (2007); (5) Eyles (2013, unpublished).

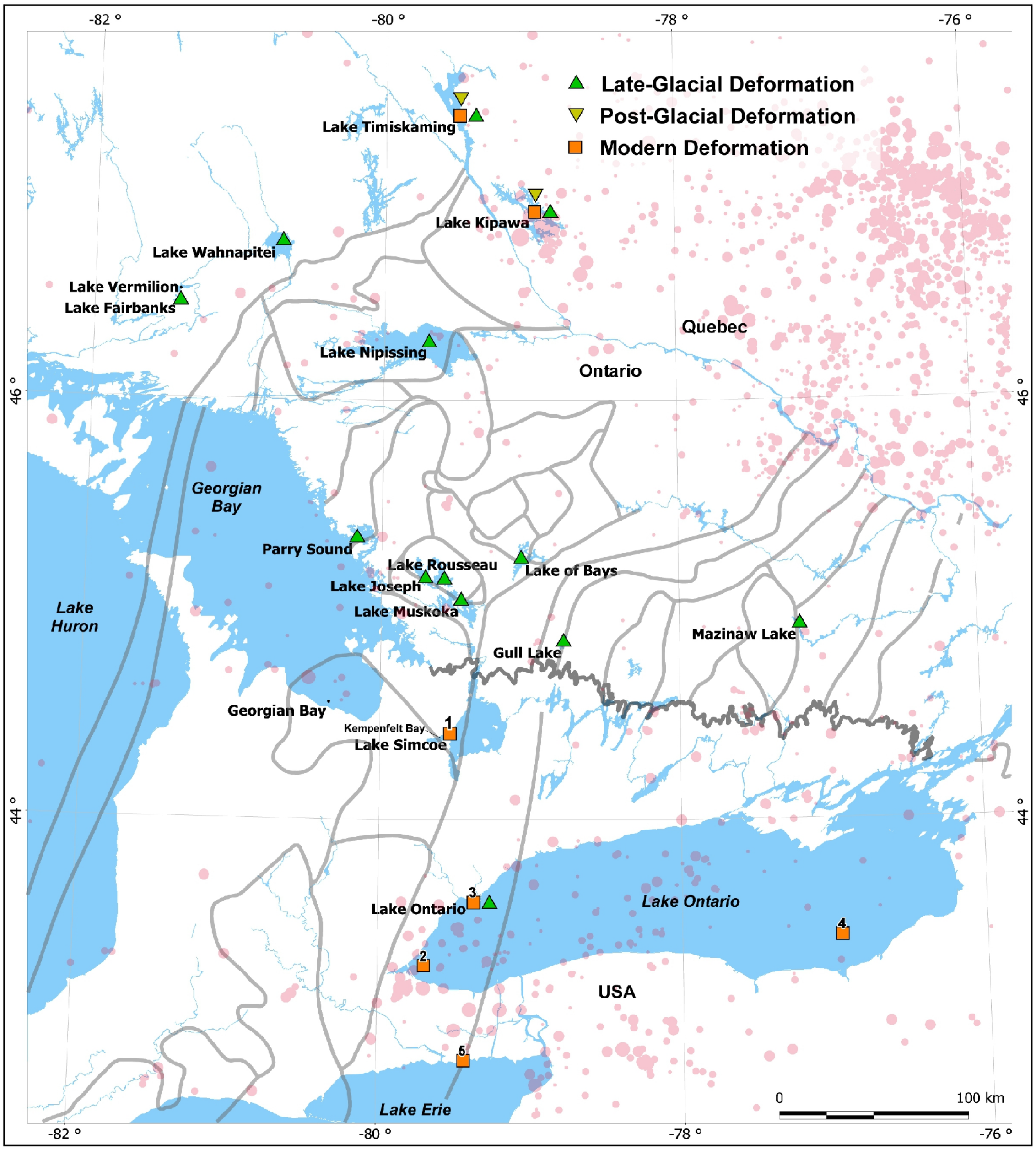

Figure 23: Continued ...

Chapter Six

6.0 Conclusions and Recommendations for Future Work

The Timiskaming Graben lies along the Ontario-Quebec border, an outlying region of the Western Quebec Seismic Zone (WQSZ; a belt of heightened intracratonic seismic activity; Forsyth, 1981) in eastern Canada. The graben itself forms part of a regional failed rift system (the St. Lawrence Rift system; SLRS) that includes the Hudson Valley in the northeastern USA, and contains several large metropolitan areas or cities within its boundary (e.g. Montreal, Ottawa, New York). The SLRS formed in response to Iapetan rifting during the breakup of Rodinia and is associated with failed rift arms below the Ottawa and Saguenay Rivers. Iapetan-age horst and graben faulting has been identified along the estuary of the St. Lawrence River between Rimouski and Baie-Comeau with some data suggesting reactivation of the SLRS. While the morphological and sedimentary record identify Holocene earthquake activity along the SLRS (e.g. Locat et al., 2003; Aylesworth et al., 2000; Syvitski and Schafer, 1996; Shilts et al., 1992), the discovery of modern faults in the Lake Timiskaming district is of considerable significance as it is one of the few locations documenting intracratonic, neotectonic faults as yet recognized anywhere in eastern North America (see also Adams et al., 1991), despite intensive study of active seismic source zones (e.g. the New Madrid; Grant, 2002; Tuttle, 2001; Johnson and Schweig, 1996).

The Timiskaming Graben forms a prominent, 50km wide, fault-bounded morphotectonic depression (resulting from subsidence within the Canadian Shield) and is partly filled by Lake Timiskaming, a post-glacial successor to the much larger glacial Lake Barlow and Lake Barlow-Ojibway. Lake Barlow-Ojibway was a regionally extensive ice-dammed glacial lake formed during retreat of the Laurentide Ice Sheet which drained approximately 8,000 ybp leaving a large, extensive clay plain. High resolution, sub-bottom and bathymetric multi-beam data collected from Lake Timiskaming reveal both late-glacial Barlow and post-glacial Holocene sediments deformed by neotectonic horst and graben structures. The bathymetry of the lake floor itself is controlled by faulting and graben basins record enhanced postglacial subsidence between parallel bounding faults. Normal faults identified on sub-bottom seismic profiles affect Lake Barlow laminated silty-clays and overlying post-glacial sediments and indicate recent neotectonic faulting consistent with the ongoing seismicity of the WQSZ (Chapters 3, 4 and 5). This faulting does not appear to be the result of the melting of buried, remnant ice blocks.

Additional seismic profiles, totaling more than 2000km line length, were collected from many other lakes in Ontario and western Quebec to constrain the record of postglacial seismicity of the region (Chapter 5). These data confirm that unconsolidated late- and post-glacial sediments deposited in lakes subsequent to deglaciation function as natural seismographs recording events in the form of co-seismic faults, slumps, debris flows and diapiric water escape structures (i.e. 'seismites'). These data also allow the discrimination of late-glacial faulting, likely related to rapid crustal rebound (as seen in Parry

Sound, Lake Muskoka and Lake Joseph) from postglacial neotectonic structures (e.g. Lake Simcoe, Boyce et al., 2002, Todd and Lewis, 1993; Lake Timiskaming and Lake Kipawa, Doughty et al., 2013, 2010a, 2010b). The largest number of post-glacial, earth-quake related deformation structures were found to occur in Lake Timiskaming (Chapters 3 and 4) and Lake Kipawa (Chapter 2) recording persistent deformation within the North American plate. Extensive slumping identified in Lake Kipawa, close to the epicentre of the 1935 M6.2 Timiskaming earthquake (Bent, 1996; Hodgson, 1936), affect both the gla-cial Lake Barlow sediments (approximately 8,000 ybp) and the overlying post-glacial gyttja. This suggests that the 1935 earthquake was the largest postglacial seismic event in this area.

Geologic and geophysical investigations of lake-sediments support the model of the Timiskaming Graben as a deforming, intraplate weak zone within the North American plate (Chapter 4). The scale and type of neotectonic deformation possibly reflect the fre-quent seismic activity and associated crustal deformation in the WQSZ. Large, damaging earthquakes have occurred within the WQSZ (e.g. Montreal, 1732 – M5.8; Timiskaming, 1935 – M6.2; Cornwall, 1944 – M5.6) with M>3 earthquakes occurring every two years in the Timiskaming district (Adams et al., 2000). The most recent large earthquake (M5.2; January 1, 2000) occurred under Lake Kipawa within 15km of the 1935 Timiskaming epicentre (Adams et al, 2000). This scale of activity is not consistent with a model of mid-plate seismicity driven by postglacial rebound alone (e.g. Chung, 2002) but, rather, with that of brittle deformation of the fractured/faulted upper crust of the North

American plate as a consequence of the interaction of plate movement and old structures (Mazzotti, 2007). Examination of regional earthquake epicentres reveals an association with major suture zones and failed rifts that record the formation and breakup, respectively, of the supercontinents Rodinia and Pangea. As such, it is hypothesized that all Precambrian structures could be considered potentially seismogenic (Chapter 5).

The detailed geophysical and sedimentological studies presented here have major societal relevance in areas of eastern North America affected by intraplate earthquakes. The recognition and mapping of earthquake related features (i.e. 'seismites') in lakes for seismic risk analysis is a means of constraining seismic recurrence intervals and more realistically assessing seismic risk across the populated area of Ontario and Quebec where events occur on time scales much longer than recorded history (Strasser et al., 2013).

6.1 Future Directions

The research presented in this thesis has emphasized the use of lake-based geophysical methods (i.e. sub-bottom reflection seismic, multi-beam bathymetry and sidescan imagery) as investigative tools for determination of lake-bottom deformation structures and their association with late- or postglacial seismicity or other processes. Future directions for research would include both local- and regional-scale analysis of additional data pertaining to the identification of past seismic events. A number of potential projects are identified below.

Lake Timiskaming - Evaluation of a (possible) fault scarp

Landslides, sand blows and/or dykes (liquefaction features), fault offsets (horizontal; strike slip) and fault scarps (vertical; dip-slip) are possible surface expressions of seismic events (Sherrod et al, 2004; Locat et al, 2003; Stewart et al, 2001; Tuttle, 2001; Vanneste et al, 2001; Russ, 1982). For example, in the Lake Timiskaming area a well-defined, linear scarp approximately 20 km long and 7 m high, can be identified on the ground surface between Thornloe and New Liskeard, and may be interpreted as a possible northern extension of one or more of the lake bottom faults found in Lake Temiskaming (Chapter 3, Figures 4 and 7).

Characterization of this scarp (i.e. the ‘New Liskeard-Thornloe scarp’) through the use of ground-penetrating radar (GPR) and seismic reflection techniques is proposed. The methods are complementary to each other with GPR imaging the shallow subsurface and seismic recording deeper reflectors. Also, as each relies on different characteristics of the subsurface (GPR reflections resulting from changes in electromagnetic parameters; seismic from changes in acoustic parameters) the two methods can be used ‘to identify interfaces across which both EM and acoustic parameters vary’ (p. 627 , Baker et al, 2001) in order to more completely characterize subsurface lithology and structure. The datasets will be processed and examined for indications of layer offsets (faults), sand dikes or sand sills (Tuttle, 2001) and deformed layers indicative of a tectonic (fault scarp) formation. A shallow borehole (or boreholes) could be drilled and logged as a check against the geophysical data in addition to trenching of the feature (Stewart et al, 2001; Vanneste et al, 2001, Tuttle, 2001). For comparison purposes, the 'Timiskaming West Shore Fault' (see Chapter 3, Figure 4A, and Chapter 4, Figure 4) should also be examined using a similar methodology.

Lake Timiskaming – Surficial features possibly related to seismicity

Geomorphological investigation of the area through the use remotely sensed data (including aerial photography) and high resolution digital elevation models can be used to determine the existence and location of any other supporting evidence for seismic activity. Surficial expression of features (other than an offset scarp) could include sand blows or dikes (Tuttle, 2001), disturbance or displacement of morphological features such as shore-lines, drainage patterns (i.e. drainage deflections) or glaciated surfaces (Stewart et al, 2001). In particular, LIDAR (light detection and ranging) can be used in the creation of high-quality DEM's (i.e. HQDEM; with ground resolutions on the order of 1-5 m) from which (potentially small-sized) features can be interpreted (e.g. faults with small throws - Begg and Mouslopoulou, 2010; historic slope failures – Jaboyedoff et al, 2012 and Van Den Eeckhaut et al, 2011; non-seismic glacial features – Smith et al, 2006).

Central Ontario – Sediment deformation and/or faulting related to bedrock structures

Surface lineaments and faults interpreted from aeromagnetic and gravity data (e.g. Ontario Geological Survey, 1999) can be used to discern regional trends related to bedrock geology and possible controls on Paleozoic (or neotectonic) faulting (Boyce and Morris, 2002; Forsyth, 1981). Aeromagnetic and gravity data can be processed after methodology developed by Boyce and Morris (2002) to 'optimize resolution of basement geophysical lineaments' (p. 159). Local magnetic data (as available) can be used to better resolve lineations seen at a regional scale as well as providing detail for identifying localized bed-rock discontinuities for more direct comparison with lake bottom deformation/disturbances (Eyles et al, 2003; Boyce and Morris, 2002). Other remotely-sensed datasets including Landsat imagery, digital elevation models (e.g. Shuttle Radar Topography Mission, Jarvis et al., 2008) and aerial photography can also be used to assist in the interpretation of surface faulting, lineaments and geomorphological features (Harris et al, 2006; Goodacre et al, 1993; Forsyth, 1981). All data and interpretations can be incorporated within a common spatial framework within a GIS such that a series of evi-dential maps (as used in spatial analysis) could be created (after Harris et al, 2006, and Daneshfar and Benn, 2002) based on lineament orientation, density, density classes and buffers. Each of these data sources provide a spatial pattern that could be related against the distribution of epicenters or interpreted deformation structures (i.e. the feature(s) of interest). If any relationship exists, it may be tied into the seismicity of the rift system. By identifying additional features that indicate the frequency and magnitude of seismic events in eastern North America, more accurate assessments of future seismic risk will be generated. This is extremely important for the future safety of large population centers in Ontario and Quebec.

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