Evidence of Soft-Sediment Post-Depositional Deformation in Lake Superior

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I never admitted the doubts I've had about finishing this paper, and without a partner who knows what's best for me, those doubts may have become certainties. My wife, Shannon, is the best teammate I could have asked for. Without her support and confidence, I wouldn't be as happy as I am today. Sorry this project took so long. I promise I'll be better about the next one.

Abstract

In the very fine-grained, gas-poor sediments of Lake Superior, it is possible to collect extremely high resolution images of the subsurface with a 28 kHz echosounder. Penetration depths routinely exceed 20 m. This represents a significant portion of the lake's soft sediment section and makes it possible to image sediments that were deposited during the most recent de-glaciation of the basin. A high-resolution "pseudo-3D" seismic survey recently collected with this system in western Lake Superior provides convincing evidence of buried iceberg scouring in the lakefloor glaciolacustrine sediments. In seismic profiles, the scour appears as localized regions of acoustical blanking that are frequently "U" or "V"-shaped. The scour zones are typically 10s of meters across and up to 6 m deep. In plan-view, they exhibit a curvilinear appearance, frequently running for several kilometers. Features associated with the iceberg scouring include thickening and thinning of beds caused by the building of lateral and terminal berms; and overlying normal faulting due to sediment collapse into the scour marks as the deformed sediment is reconsolidated. Ring-shaped depressions on the modern lake floor are theorized to be the result of sediment dewatering linked to the development of the underlying features.

Table of Contents

List of Figures

List of Tables

Chapter 1: Introduction

The Lake Superior basin has undergone a complex evolution throughout the Quaternary Period as a result of repeated glacial advances and retreats. Moving along the NE-SW axis of the lake basin, the Superior Lobe of the Laurentide Ice Sheet (LIS) left the basin about 10,000 years ago. This glacial history is recorded in the mineralogy, stratigraphy, morphology and structure of the sediments in and around Lake Superior.

Past work in the area studied the presence and origin of curious ring-like topographic features (Figure 1.1) on the lake floor using high-resolution seismic data (Figure 1.2) (Flood & Johnson, 1984; Wattrus, et al. 2003; Wattrus, 2004). Analysis of sub-surface sediment characteristics revealed several interesting structural and sedimentological features that may be related to the unusual lake floor rings. Through the use of seismic and sediment core data, this research seeks to identify the origin of these features and characterize them as evidence of glacial and peri-glacial lacustrine post-depositional sediment remobilization in Lake Superior.

In modern environments the volume of sediment influx, lake currents, erosion, post-depositional processes, biological activity and changes in climate affect the distribution and related structures of lake sediments. When studying environments of glacial deposition, processes such as ploughing, iceberg scouring, isostatic rebound, and sediment dewaterting must be considered. In the specific case of Lake Superior's glacial history, the interrelationship between the Laurentide Ice Sheet and Glacial Lake Agassiz also needs to be taken into account.

Through the analysis of the high-resolution seismic record and sediment cores, this study will provide evidence to answer four general questions:

- Has post-depositional remobilization of the glacial sediments occurred?

Specific features represented in the seismic record lead us to believe the sediments have been disturbed after their initial deposition. These features correlate directly to physical evidence of remobilization within the sediment cores taken from the study area.

- How do these remobilization events fit into the established glacial history of the Lake Superior basin?

Using what is known about the geologic history of the study area, this study will assign a chronological order to the seismic features. Inferences can be made regarding how the ice sheet, the lake basin, and other regional glacial features (e.g. Glacial Lake Agassiz) played a part in the remobilization.

-What is the mechanism for the post-depositional remobilization?

A plausible mechanism that led to the formation of the features will be presented.

-Has the remobilization of the sediments influenced the development of the ring-like structures on the floor of Lake Superior?

There are distinct connections between features observed in the seismic section and the overlying ring structures. A possible connection between the features will be presented.

Chapter 2 begins with a general summary of past research conducted in the Lake Superior basin. I provide general information about the physical characteristics of this study area as well as details about the bedrock geology and Quaternary history. This will include essential facts about the Laurentide Ice Sheet and Glacial Lake Agassiz that are important to my interpretation of Lake Superior's sediment record. The final portion of the chapter will focus on post-glacial events that influence the basin evolution.

Chapter 3 is an outline of the methods and equipment used to collect both seismic and core data. I will review the processing of the data and explain the why both seismic and core data are required for this study. My interpretation of this data is laid out in chapter 4. I discuss the characteristics and distribution of seismic facies and describe the core samples and notable features within the sediment record. I will provide a correlation between the data sets that make it possible to see the study in a "bigger picture" context.

I will outline past hypotheses for deformation of the soft-sediment record in Lake Superior in chapter 5: Discussion; and integrate the previous chapters' information to present an explanation of the events that have led to the post-depositional remobilization of the sediments. The summary of this thesis in chapter 6 recaps the key results and emphasizes the impact that the Laurentide Ice Sheet and Glacial Lake Agassiz had on the development of Lake Superior's sediment record.

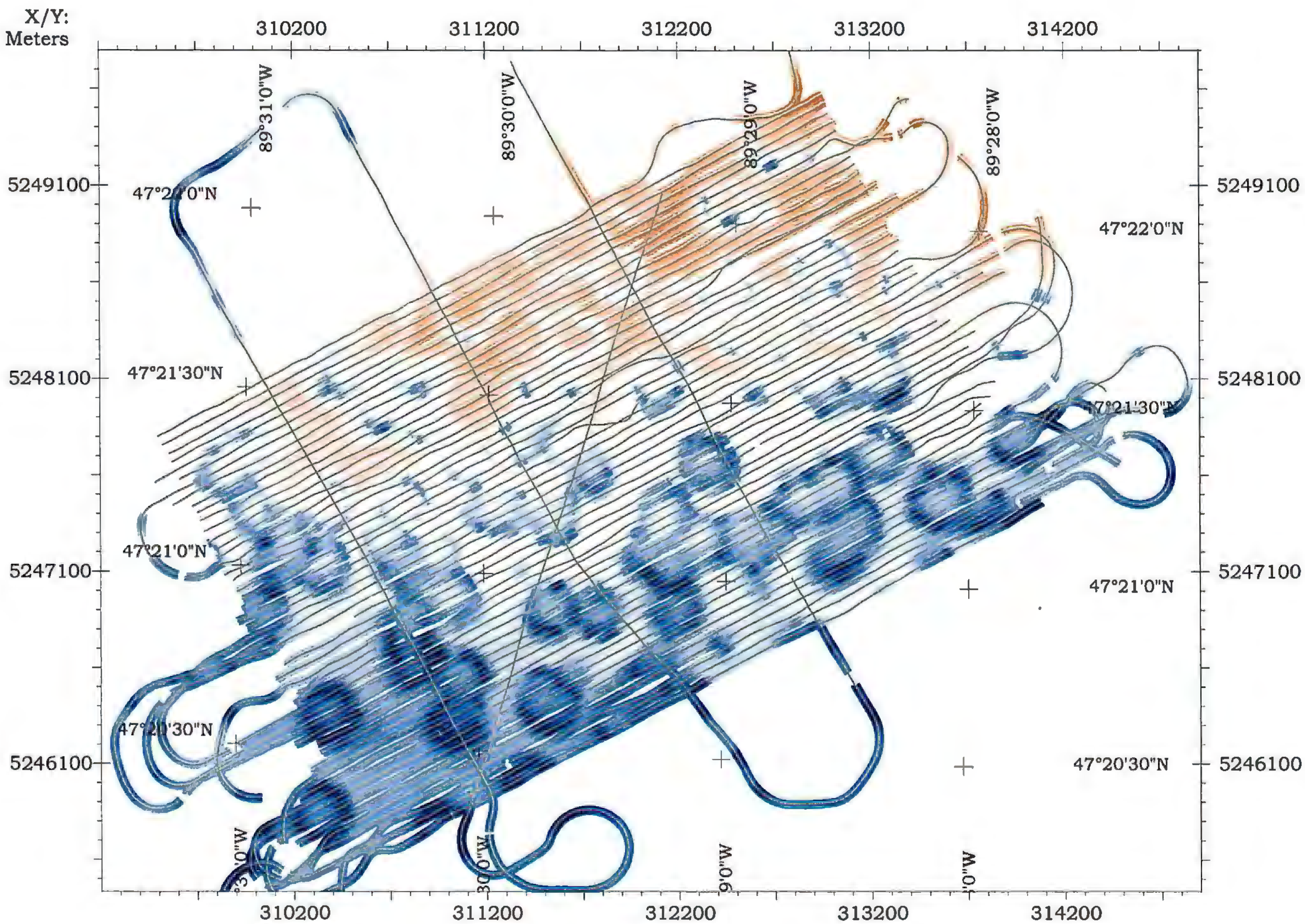


Figure 1.1: Seismic representation of lake floor in study area showing ring-like morphology.

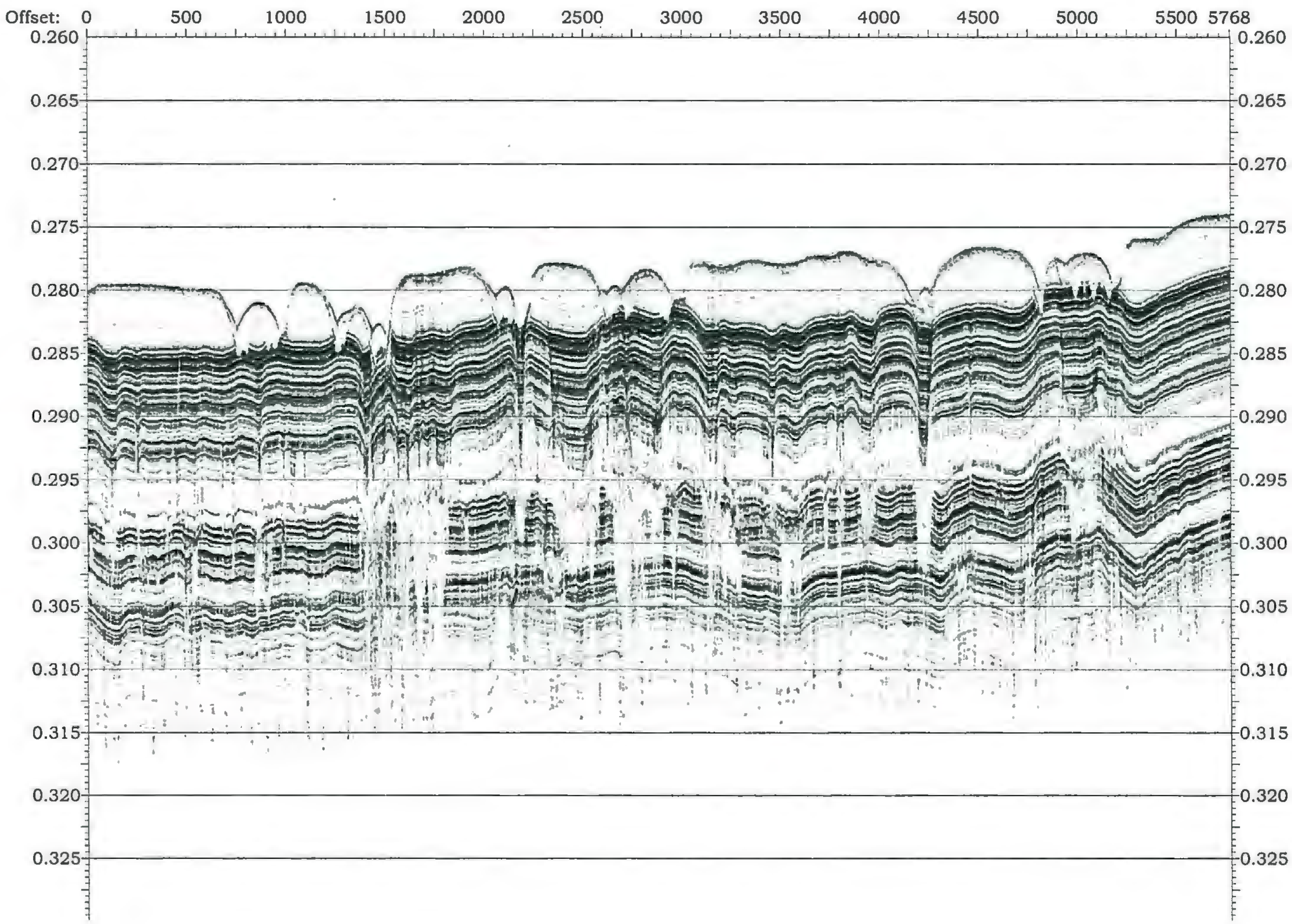


Figure 1.2: Example of sub-bottom seismic data taken from the study area.

Chapter 2: Background

2.1 History of the Study of Lake Superior

Louis Agassiz published his theory of ice ages in *Étude sur les glaciers* in 1840, and made observations of ancient lake terraces as well as the flora and fauna of Lake Superior as early as 1850. Detailed studies of the shorelines occurred in 1893 by A.C. Lawson, and again by F.B. Taylor in 1895 and 1897. Lawson believed the shorelines to have remained horizontal, Taylor disagreed.

From the 1930's until the late 1950's study continued to focus on the relict shorelines, with an emphasis on making correlations across the basin to provide an understanding of the evolution of the lake. In 1961, Zumberge and Gast explored the nature and thickness of the glacial sediments by coring multiple locations within the lake basin. In 1965, S.C. Zoltai used radiocarbon dates and mapped glacial deposits and lake sediments to significantly narrow down the northern and eastern extents of Lake Agassiz and Lake Superior. Studies of the basin's lake evolution and drainage history were published in 1981 by Futyma and Drexler. Analysis of this previous work and further study of the interrelationship of the Agassiz and Superior basins were published in 1983 by Teller and Thorliefson.

2.2 Lake Superior Physiography

Lake Superior is located at the southern edge of the North American Precambrian Shield and is the largest of the present day Great Lakes. It is approximately 725 km long, 280 km wide and has an average depth of 150 meters and maximum depth of 406 meters (Figure 2.2.1) (Johnson, 1980 and Wattrus, 2001). Lake Superior's drainage basin encompasses about 210,000 square km, utilizing some 200 tributary rivers (Matheson and Munawar, 1978; Landmesser et al. 1982). Present day Lake Superior is the product of glacial erosion. The general basin morphology exhibits three distinct regions, a relatively flat, sloping western margin; a wide, gentle trough in the center of the basin; and a series of north-south trending ridges and valleys to the eastern end of the basin (Wold et al., 1982).

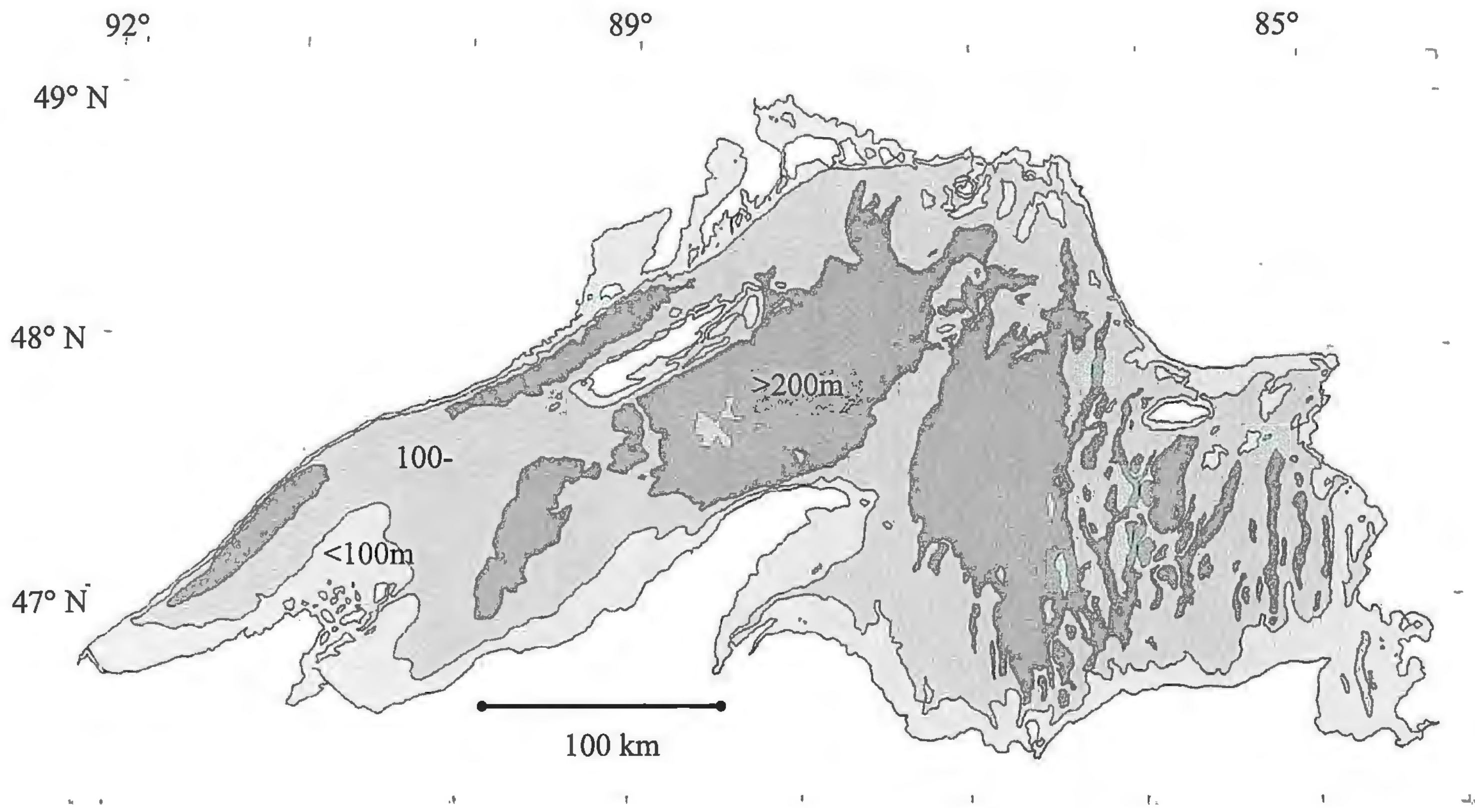


Figure 2.2.1 Lake Superior bathymetry and basin morphology.

2.3 Bedrock Geology

The Superior Province contains ancient rocks, some dated to 2.7 Ga. There are three sub-provinces in the Superior; the Wabigoon, Wawa, and the Quetico. The Wabigoon and Wawa sub-provinces are composed of volcanic rocks of varied composition, and are thought to be the result of the collision of multiple island arcs. Metasedimentary shale and greywacke are also present in the Wabigoon and Wawa. The Quetico sub-province is made up of metasedimentary lithologies formed as an accretionary wedge along a subduction zone. The Quetico is mainly high grade schist, and gneiss (Card, 1990; Goodwin 1991).

Around 1.9 Ga conglomerate, sandstone, shale and banded iron formation associated with the Penokean Orogeny were deposited in foreland basins to the south and west of the current Lake Superior basin (Klasner et al., 1991). To the north are 1.4 billion year old sandstone, dolomite, shale, and evaporite of the Sibley Group.

The rocks surrounding the Lake Superior basin are generally known as the Keweenawan Supergroup, which includes the Duluth Complex of anorthosite and gabbro, the North Shore Volcanic Group of basalt and rhyolite, and the post-volcanic sediments of the Oronto Group and Jacobsville Sandstone/Bayfield Group (Figure 2.3.1). These units are predominately the result of plutonic and volcanic activity during the opening of the Mid-Continent Rift (approximately 1.1 Ga). The rift extends 2000 km from central Michigan north to Lake Superior, where it turns southwest, following the trend of the Lake's north shore through Minnesota, Iowa, Nebraska, and Kansas. Post-volcanic clastics of the Keweenawan Supergroup are mainly conglomerate, sandstone, and shale that were deposited by fluvial processes and may be up to 2100 m thick (Green, 1982; Ojakangas and Morey, 1982). Seismic investigations and drill cores indicate that the majority of bedrock under the lake floor is composed of the Jacobsville sandstone (Farrand, 1969).

The main structural feature of the western Superior lake basin is the Lake Superior Syncline (Figure 2.3.2), which extends roughly northeast from the Bayfield Peninsula of northern Wisconsin. The presence of the syncline, along with the exposure of readily eroded post-volcanic sediments probably determined the location of the basin (Manson and Halls, 1997; Wold et al., 1982). Normal faults named: Isle Royale, Douglas, Keweenaw, Thiel, and Michipocoten are present in the basin and are the result of the extension associated with the Mid-Continent Rift. The faults were reactivated with reverse movement during compression that followed the rift phases (Manson and Halls, 1991).

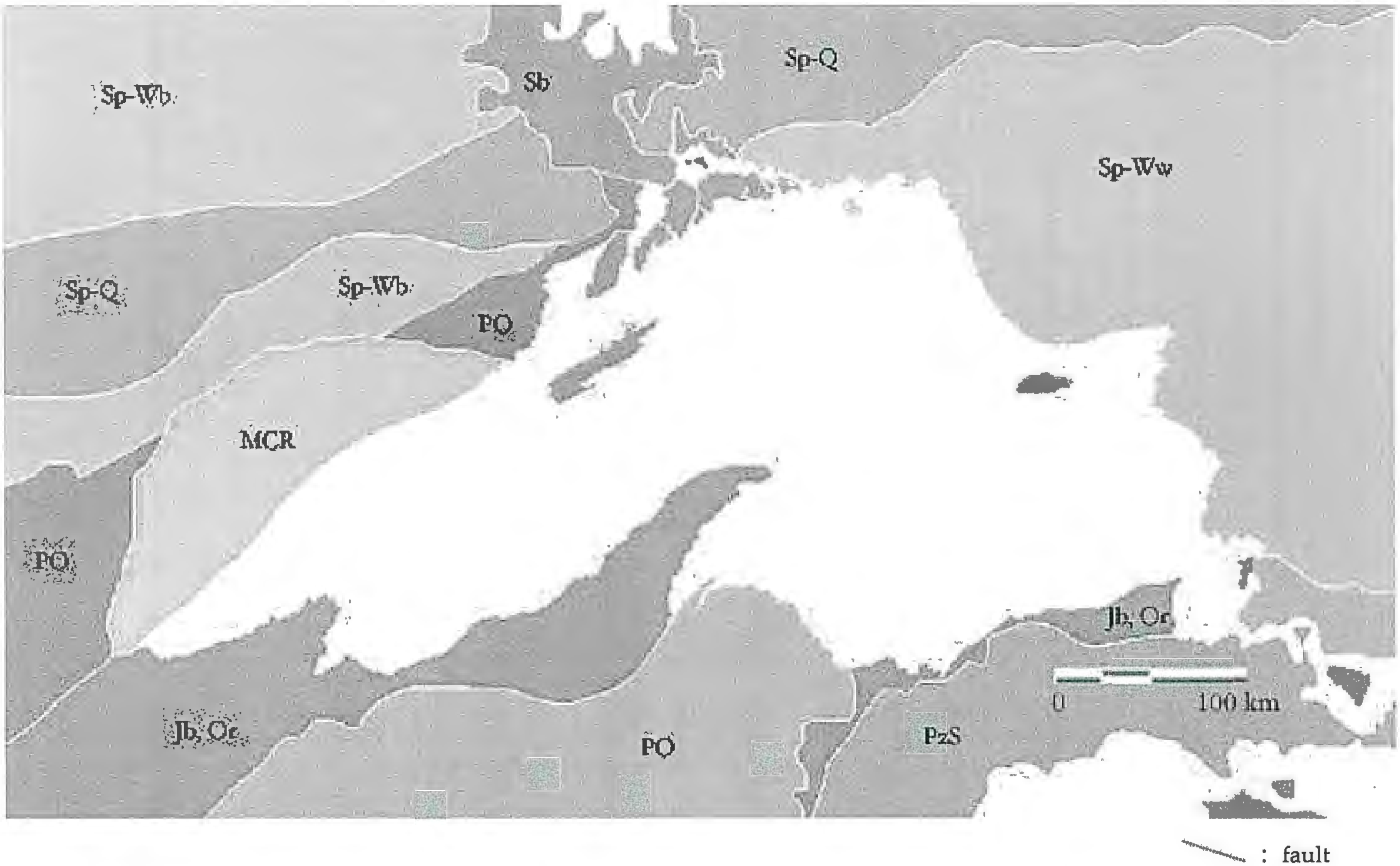


Figure 2.3.1 Geologic and Structural Map of Lake Superior (after Breckenridge, 2005; Green, 1982; and Huber, 1975). Sp-Wb=Superior Province-Wabigoon; Sp-Q=Superior Province-Quetico; Sp-Ww=Superior Province-Wawa; Sb=Sibley Group; PO=Penokean Orogeny metasediments; MCR=Mid-Continent Rift volcanics; Jb,Or=Jacobsville and Oronto groups of metased. and volcanics; PzS=Paleozoic sandstones and shales.

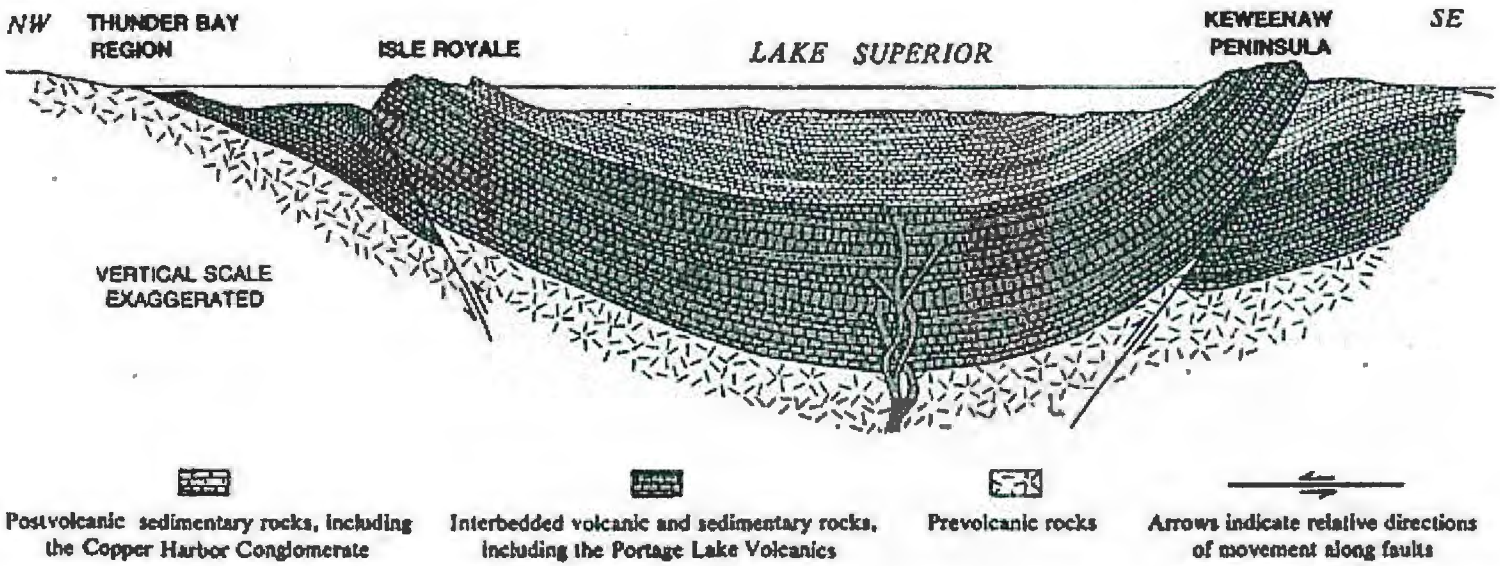


Figure 2.3.2 Cross section of Lake Superior Syncline through Isle Royale
(from Huber, 1975)

2.4 Quaternary History

The recent history of the Lake Superior Basin is defined by the interactions of the Laurentide Ice Sheet (LIS), meltwater and lakes, isostatic rebound, and changes in regional and global climate (Breckenridge, 2005). Quaternary sediments in the lake lie unconformably on Precambrian basement rocks and are the product of multiple cycles of advance and retreat of the Superior lobe of the LIS. The majority of these sediments are the result of deposition since the last glacial maximum, the late Wisconsinan. The sedimentary record of glacial activity prior to this time has been removed or reworked (Zumberge & Gast, 1961; Farrand, 1969).

2.4.1 Glaciation

The Superior lobe of the LIS advanced from the northeast into the region along the western axis of the Superior basin. The advance is recorded in the sedimentary record by a layer of red sandy and clayey till that is characterized by a distinctive clast provenance consisting of basalt, agate, felsites and sandstones (Wright et al., 1973). Five major advances of the Superior lobe are documented in Minnesota. These include: the St. Croix phase (oldest), which reached its maximum in southeastern Minnesota about ~15.5 ka BP and the Automba phase, which flowed to the north and west of the previous advance (Mooers, 1988). Following the Automba phase, the glacial sediments changed from sandy to more clay-rich till, reflecting the ice sheet's retreat into the Superior basin and the formation of Glacial Lake Duluth. During the next phase, the Split Rock, the Superior lobe entrained the lake basin clays and deposited them as the Cloquet moraine ~13 ka BP.. The last advance of the Superior lobe outside of the Superior basin was the Nickerson phase, ~12 ka BP, producing the Nickerson and Thompson moraines to the southwest of the present lake basin (Figure 2.4.1) (Wright et al., 1973). The discovery of the Gribben Lake buried forest provided radiocarbon dates that identified the final advance of the LIS within the lake basin ~10 ka BP. Named the Marquette phase, the final position of this advance is debated, however Clayton (1983) stated the margin advanced "nearly to Duluth."

As the Superior lobe retreated northward out of the lake basin, several proglacial lakes occupied the region (Carney, 1996). These ultimately combined into Glacial Lake Duluth, which was drained through the St. Croix River in Minnesota and the Brule River in Wisconsin. Evidence of the lake is preserved in various beach levels surrounding the basin at around 330 meters a.s.l. (Farrand, 1969). As the Superior lobe continued to retreat, the exposure of lower outlets led to progressively lower lake levels in the proglacial lake. Drainage of Glacial Lake Duluth shifted from the Brule and St. Croix drainages between 330 and 315 m a.s.l. to the Powell lake channel at 305 m a.s.l., the lake then began a 1000 year progression of lower post-Lake Duluth levels.

With the continued retreat of the ice, the water level in the Superior basin dropped to 43 meters below current lake level during the Lake Minong stage, when it was drained by the St. Mary's river (Farrand, 1969). The lowest level of the lake occurred when the ice retreated far enough to allow the water to drain at North Bay, Ontario, 69 m below current lake level. Lake level began to rise again as the North Bay outlet began to rise due to isostatic rebound with the unloading of the glacial ice. The current lake level was achieved approximately 5000 years ago as Lake Nipissing. Rebound eventually diverted the outlet of the lake to Port Huron 4000 years ago, and at 2000 years BP, rebound caused the outlet to change to the present sill at Sault Ste. Marie, where Lake Superior has settled at its current level of 183 meters a.s.l. (Figure 2.4.2; Farrand and Drexler, 1989).

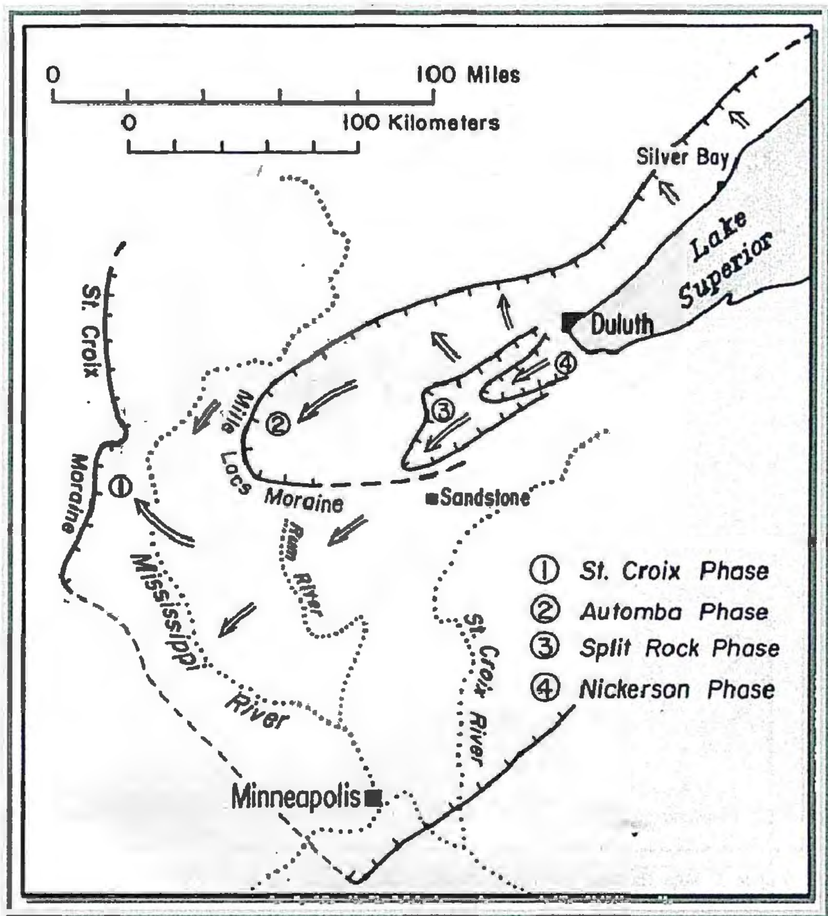


Figure 2.4.1 Four phases of Superior Lobe retreat (from: Wright, 1973). Tunnel Valleys, Glacial Surges, and Subglacial Hydrology of the Superior Lobe, Minnesota, *in* Black, Robert F., Goldthwait, Richard, Willman, H.B., eds., *The Wisconsinan Stage*. Boulder, CO: The Geological Society of America, Memoir 136, pg. 253.

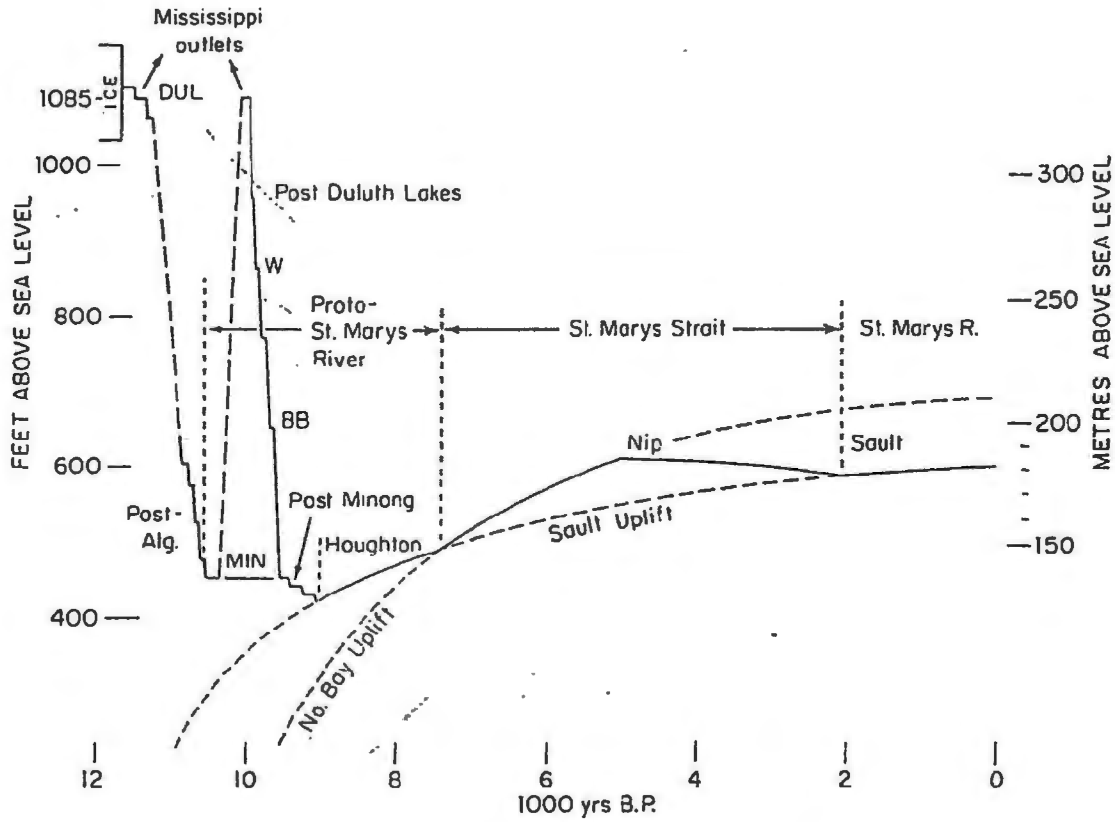


Figure 2.4.2 Water level curves in the Lake Superior Basin prior to isostatic rebound. The spike at 10 ka B.P. occurred only in the western basin, while the eastern basin remained at the Minong level. DUL=Duluth, POST-ALG=Post-Algonquin, MIN=Minong, W=Washburn, BB=Beaver Bay, NIP=Nipissing. (from Farrand and Drexler, 1985)

2.4.2 Glacial Lake Agassiz

Lake Agassiz was a vast ice-marginal glacial lake that occupied parts of Saskatchewan, Manitoba, North Dakota, Minnesota and South Dakota at various times in its existence (Teller, et al., 2002). The lake formed in the Hudson Bay drainage between 13 and 8 ka BP. The lake basin was the collection area for over 2,000,000 square km of land as well as Laurentide Ice Sheet meltwater.

Lake level rose and fell multiple times during the history of the lake in response to changes in the position of the margin of the LIS and outlet elevation (Teller et al., 2002; Teller & Leverington, 2004). Drainage occurred through outlets to the northwest, south, and east and later to the north (Figure 2.4.3). The eastern and northwestern outlets of the lake were controlled by ice dams. When these outlets were opened due to ice retreat, draw-down of the lake to the new drainage elevation was dramatic. Table 1 lists the various stages of Lake Agassiz, as well as estimates of hydrology and what direction the outlets drained.

LAKE STAGE	LAKE VOLUME km^3	LAKE DROP m	OUTBURST VOLUME Km^3	FLOOD FLUX Sv	ROUTE OF OUTBURST	AGE OF OUTBURST ka BP
HERMAN	10900	110	9500	0.30	EAST	12.9
NORCROSS	13300	52	9300	0.29	SOUTH, NW	11.7
TINTAH	18555	30	5900	0.19	SOUTH, NW	11.2
UPPER CAMPBELL	22700	30	7000	0.22	EAST	10.6
LOWER CAMPBELL	19100	16	3700	0.12	EAST	10.4
McCAULEYVILLE	16400	10	2100	0.07	EAST	10.3
HILLSBORO	19200	7	1600	0.05	EAST	10.0
BURNSIDE	10300	12	2300	0.07	EAST	9.5
THE PAS	4600	12	1600	0.05	EAST	9.2
KINOJEVIS	163000	770	163000	5.2	NORTH	8.4

Table 1; Lake Agassiz Stages (from Teller, et al. 2002)

Between 10,600 yr BP and 9,200 yr BP, several of the drainage events occurred via an eastern route of the Lake Superior basin and the St. Lawrence Seaway to the North Atlantic. Based on estimated paleogeometry of overflow channels, and depth of flow, it is inferred that the outburst events could have occurred over a period of "a few months to a few years, in all cases probably less than a decade" (Teller, et al., 2002).

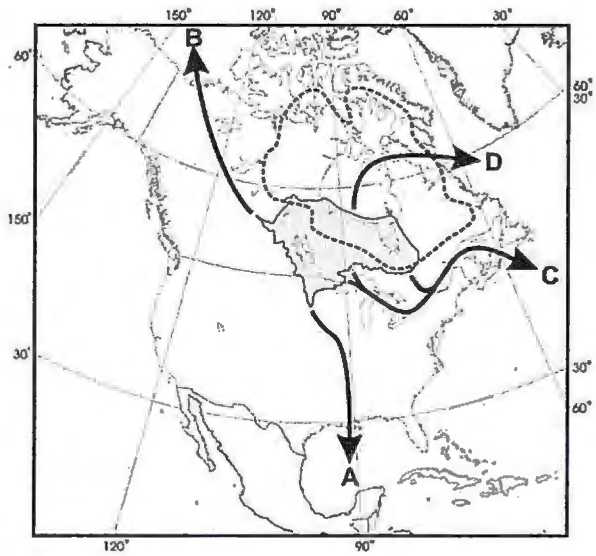


Figure 2.4.3 Drainage routes (A-D) of Glacial Lake Agassiz. General outline of LIS at 10 ka BP shown with dashed line (from Teller et al., 2001).

2.4.3 Climatic Significance of Glacial Lake Agassiz

The Younger Dryas cooling event, ca. 13,000-11,400 cal. yr. B.P. has been linked to a reduction in the rate of formation of North Atlantic Deep Water (NADW) caused by an increase in the freshwater flux to the North Atlantic (Broecker et al., 1988, 1990; Clark et al., 2001). Fluctuations of the LIS during a period of general recession modulated this freshwater flux. For example, ice sheet retreat opened an eastern Lake Agassiz outlet into Lake Superior. This allowed large volumes of freshwater to flow to the Atlantic via the Great Lakes and the St. Lawrence River. Readvance of the ice to the Marquette/Grand Marais/Munising position closed this outlet, temporarily returning Lake Agassiz flow to the Gulf of Mexico by way of the Minnesota and Mississippi rivers (Glacial River Warren) (Teller and Thorleifson, 1987).

Additional Lake Agassiz floods associated with the most recent deglaciation of North America are thought to have been large enough to disrupt the formation of NADW, perhaps initiating the Pre-Boreal Oscillation (PBO), ca. 11,200-11,050 cal. yr. B.P., and the "8.2 cal ka event." Catastrophic drawdown of Lake Agassiz into the Mackenzie River basin has been linked to the PBO (Fisher et al., 2002; Teller et al., 2002), and the 8.2 cal. ka event is attributed to final drainage of Lake Agassiz via the Hudson Strait (Barber et al., 1999; Teller et al., 2002; Teller and Leverington, 2003; Clark et al., 2003, 2004).

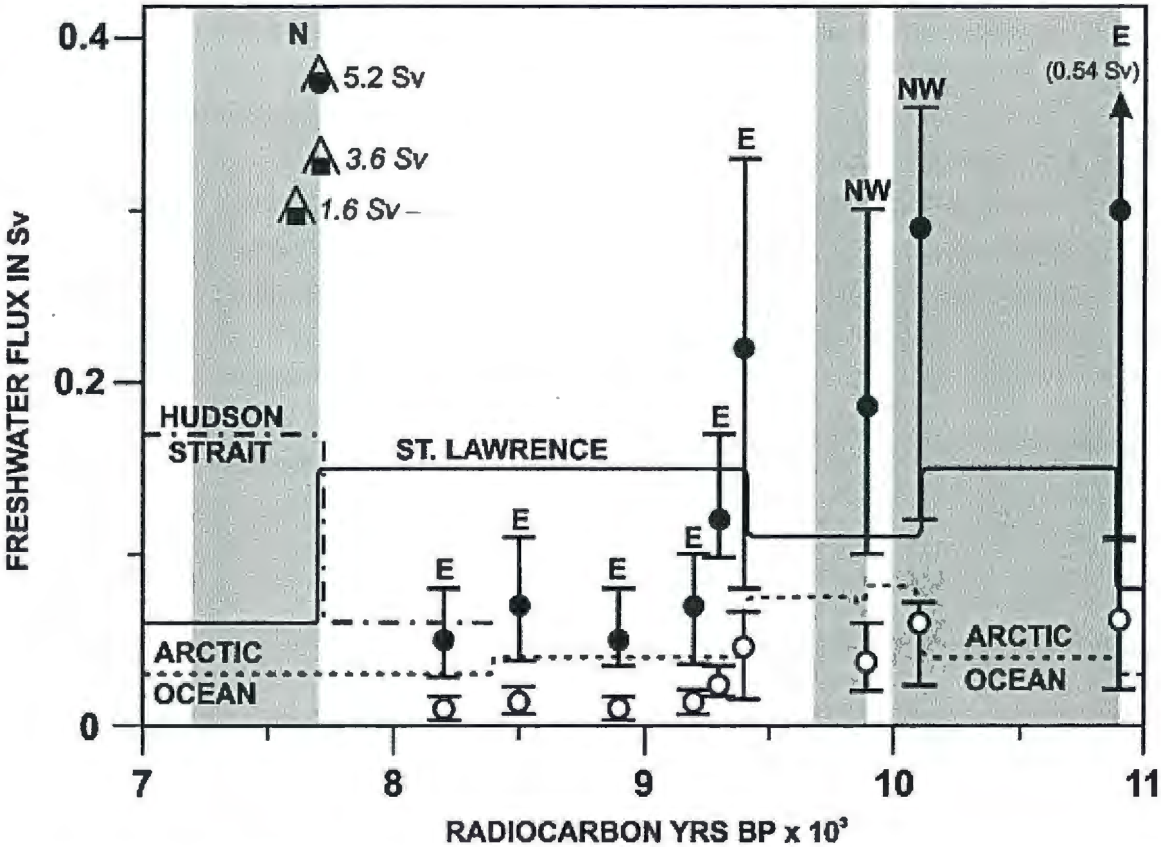


Fig. 2.4.4 Flux of 10 outbursts (in Sv) from Lake Agassiz. Solid dots show flux if lake drawdown occurred in 1 year (note that values at 7.7 and 7.6 ka exceed scale) and open circles show flux if drawdown was in five years; the two squares at 7.7–7.6 ka are alternative fluxes with a two-step draw down. The 1-year outbursts are those given in Table 1 as Flood Flux. Alternative ice margin positions would expand or reduce the volume of stored (and released) water; our range of flux resulting from a ±1° latitude shift of the ice margin is shown by vertical lines, except at 10.9 ka where the value exceeds the scale. Baseline flows are shown as solid or broken lines and are total fluxes of outflow through each drainage basin, including Lake Agassiz baseline flow (after Licciardi et al., 1999). St. Lawrence Valley to the North Atlantic Ocean; N=Hudson Bay–Hudson Strait to Labrador Sea and North Atlantic Ocean. The Younger Dryas, PBO, and the cold event at 7.7–7.2 ka ^{14}C yrs (the "8.2 ka cal yr cold event") are shaded. (From Teller et al., 2002)

Additional overflow events occurred in Lake Superior between 10,600 and 9,100 cal. yr. B.P., when subsequent ice margin retreat reopened eastern outlets, permitting a return of meltwater flow to the North Atlantic via the Great Lakes (Leverington and Teller, 2003). None of these latter drawdown events through

Lake Superior have been linked to abrupt cooling events (like the Younger Dryas or the 8.2 cal ka event), but two distinct reductions in the formation of NADW occur during this time period (Bond et al., 1997).

2.4.4 Soft Sediment Record

The sedimentary record of the Lake Superior Basin is a tangible resource to understand the glacial history of the region. It ties together LIS retreat and advance, and the associated intervening glacial lakes, including their levels and drainage patterns.

A drilling program undertaken in 1961 and 1962 by the University of Minnesota and the University of Michigan collected 11 cores, which penetrated up to 200m of glacial and glacio-lacustrine sediments. Four of the cores contacted bedrock. In 1966, further coring throughout the lake recovered 90 piston cores of up to 6m of sediment depth (Farrand, 1969). The portion of the record that has been sampled by seismic and coring methods for this study is restricted to the upper 8-9m of the Quaternary section.

Bedrock sampled by the '61-'62 cores was a mottled red and white sandstone interpreted as Precambrian Jacobsville sandstone. Fragments of this sandstone and well-sorted brown sands appear in the overlying glacial sediments of red clayey or silty till. The till transitions upward into stratified red clay with some silt and fine sand. This clay is superceded by varved red clays, then grey varved clays. Above the varved sequence, the grey clay becomes homogenous (Figure 2.4.5) (Farrand, 1969).

Glacial Sediments

The unconsolidated sediments that were directly deposited on top of bedrock consist of red silty to clayey till with few erratic inclusions of basalt, sandstone, gneiss and gabbro. These till layers are between 12 and 60m thick and alternate with layers of red lacustrine clay.

Fine brown sand is sometimes associated with the till. The sand is well sorted, sub-angular to sub-rounded, but can be well rounded and frosted. The sedimentary characteristics, its deep-water location within the basin, and the fact that the sand is deposited at the same stratigraphic position as the till, indicates it is an outwash sand deposited at the margin of the retreating Superior Lobe.

Lacustrine Sediments

Immediately above the red till, fine grained lacustrine silt and clays are prevalent throughout the basin, ranging in thickness from 0 to 40 meters. Locations where coring recovered no lacustrine sediments had water depths less than that of storm base, or in areas subject to continued erosion.

The lowermost lacustrine sediments are thick intermediately stratified red clays with occasional interbeds of fine sand or silt. In areas where the stratification of this layer is more developed, the layers are thick – greater than 8 cm each. The overall thickness of this layer ranges from 1 to 30m.

As the glacier retreated further out of the basin, these red clays transition upward into red varves. The lower varves are up to 4cm thick, but become thinner (1.5cm) and finer grained upwards. The combined thickness of the red varves is between 91 and 300cm. Lithologically, the red lacustrine clays, including the varves is identical to the clay included in the underlying till (Farrand 1969).

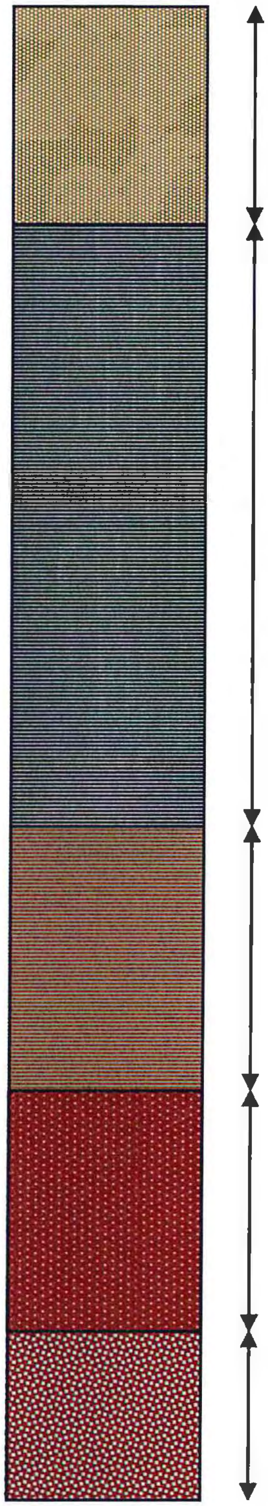
Approximately 1300 grey varves supercede the red varves (Breckenridge, 2005). The transition is gradual, with red and grey varve thickness and grain size virtually identical. Over a thickness of 30 to 100 cm, the red color changes completely to grey. The grey varves are composed of the finest grain size of all sediments recovered within Lake Superior and average 450 cm in total thickness (Farrand 1969). The grey varves stop abruptly and are capped by homogenous grey clay. The character of the sediment is nearly identical to the underlying varves, but is slightly coarser. Thickness of this uppermost section ranges from 1 to 4.5m, but is generally less than 2m thick. In shallower and western portions of the lake, a brown clay veneer caps the homogenous grey clay.

The change from red to gray sediment was originally thought to be the result of a change in oxidation state. In 1965, Nussman proposed several geochemical lines of evidence to contradict this; Eh and pH values are similar in both sediments, there is a lack of organic material anywhere in the sequence and ferric oxides in the sediments are in stable states. Instead of a change in chemistry, a change in source area is a more likely reason for the color change. As the ice retreated, the sediment provenance shifted from the iron-rich red deposits of the south and western portions of the area to grey deposits north and east of a line between Sault Ste. Marie and Port Arthur, where Precambrian shield rocks are light colored granites (Farrand, 1969).

Sedimentation rates changed significantly between the retreat of the LIS and the current ice-free basin. Deposition of sediments varies with currents, depth, and proximity to source. During the retreat, Farrand (1969) estimated the rate of deposition of glacial sediment to be more than 8 cm/yr, with that number decreasing to 1.55 cm/yr and 0.5 cm/yr within the red varved and grey varved section sections, respectively (Table 2). Current estimates of sedimentation in Lake Superior are about 0.03cm annually (Johnson, 1979).

Sediment Character	*Thickness Range*	*Sedimentation Rate*
Homogenous gray clay	10-350 cm	0.03 cm/yr
Gray varved clay	257-900 cm	0.5 cm/yr
Red varved clay	91-300 cm	1.55 cm/yr
Red clay and silt	>580 cm	8.25 cm/yr

Table 2: Lacustrine Sediment Thickness and Sedimentation Rates (Farrand, 1969)



Homogenous grey-brown clay. Faint laminations may be present. This clay unit may be absent in water depths less than 75 m. Thickness range is 7-450 cm.

Alternating light and dark grey silty-clay to clay varves. Varve couplets average 5 mm thick. Section may contain as many as 1300 couplets. Thickness range of grey varves is 257-900 cm.

Dark red/reddish brown varved sandy-silt to silty-clay. Light colored laminae are typically coarser grained, and may include ice rafted pebbles. Upper contact with grey varves is transitional over 100 varves. Thickness range of red varves is 91-300 cm.

Red clay interbedded with silt and very fine sand. Thickness range is 0 to 40 m.

Red clayey or silty till associated with Marquette advance. There are erratic pebbles composed of basalt, sandstone, gneiss and gabbro. Thickness range is 12-60m.

Figure 2.4.5 Generalized stratigraphic section of Lake Superior Sediments (modified from Breckenridge, 2005)

2.5 Lake Superior's Rings

Seismic reflection surveys have noted the floor Lake Superior is not a featureless expanse. In 1973, Berkson and Clay first observed rings or sub-circular depressions in the fine grained sediments from side-scan sonar records. They interpreted the depressions as polygonal cracks or fractures. In 1984, Flood and Johnson also noted the rings in side-scan sonar data. They described the rings as having troughs up to 5 m deep, and a lake bed depth inside the ring the same as outside the ring. Off the Keweenaw Peninsula, the rings have diameters between 100 and 300 m and trough widths between 20 and 50 m. Similar features have been observed in Lake Michigan (Berkson et al., 1975; Coleman et al., 1994) and Lake Huron (Moore et al., 1994).

The formation of the rings has been debated as the product of syneresis of the fine grained sediments, as dewatering structures due to sediment compaction, and as a system of immature, small scale, layer-bound extensional faults, called a polygonal fault system (PFS). The rings commonly occur in very fine grained sediments. They often appear to be partially infilled by current transported sediments (Flood and Johnson, 1984), suggesting formation in the past or a slow present day growth rate. The deformation of sediments associated with rings is limited to the uppermost Holocene surface sediments, and sometimes to the uppermost glaciolacustrine sediments. This suggests that the rings represent sites of erosion or non-deposition of Holocene sediments.

Chapter 3: Methods

3.1 Seismic Data Acquisition and Processing

The high resolution seismic experiment acquired in 2001 that is used for this thesis was designed to study the nature and origin of the large ring-shaped depressions that are widely developed over much of the deep lake floor in Lake Superior and their relationship to sub-surface sediments and structures (Flood and Johnson, 1984; Wattrus et al., 2003; Cartwright et al., 2004). Regional data had shown that this area exhibited well developed lake floor rings and related subsurface deformation structures.

Seismic profiles were acquired with the Knudsen 320B 28 kHz echosounder mounted to the hull of the R/V Blue Heron. An echosounder sends an acoustic pulse directly downwards to the seabed and records the returned echo. The sound pulse is generated by a transducer that emits an acoustic pulse and then "listens" for the return signal. The time for the signal to return is recorded and converted to a depth measurement by calculating the speed of sound in water. As the speed of sound in water is around 1,500 meters per second, the time interval, measured in milliseconds, between the pulse being transmitted and the echo being received, allows bottom depth and sub-bottom targets to be measured. Prior experience with this system demonstrated that in deep water areas of the lake, where the sediments are composed of very fine-grained clays, this system is capable of achieving over 20m of sub-bottom penetration at extremely high resolution. This sediment depth typically spans the period from, and including, the last deglaciation of the basin.

Data was collected by sailing a series of parallel lines, spaced at 50m intervals. The repetition rate of the echosounder is determined by the water depth. Given the survey speed of 6km/h, the average data density was 1110 records/km.

The surveyed area is approximately 4km by 2.5km. Water depth in the survey area shallows to the northwest from a depth of 215m to 200m. The lake floor rings in the area vary between 100 and 300m in diameter. Ring troughs are 20 to 30m wide and up to 5m deep. They occur both as isolated structures and in clusters. The objective of this survey design was to produce a high resolution seismic dataset with which to map the lake floor rings and the deformation structures in the underlying Holocene section. Navigation and survey control was obtained using the R/V Blue Heron's multibeam management system.

Post-acquisition processing of the data was performed using *SU*, a seismic processing package distributed by the Colorado School of Mines over the internet (Stockwell, 1997). Processing consisted of formatting to the industry standard SEGY format, editing, filtering, and gain correction. The processed data was loaded onto a workstation for interpretation using the *Kingdom Suite* Interpretation System (Seismic MicroTechnology, Houston, TX).

The reflectors were assigned a color designation and traced laterally in each line of data to correlate each seismic line across the study area. Because it is possible to identify a range of three-dimensional shapes with a closely-spaced framework of two-dimensional seismic reflection profiles; this correlation provides the "pseudo-3D" representation of the lake floor and subsurface seismic reflectors (sediment horizons) (Figure 3.2.1 through 3.2.3).

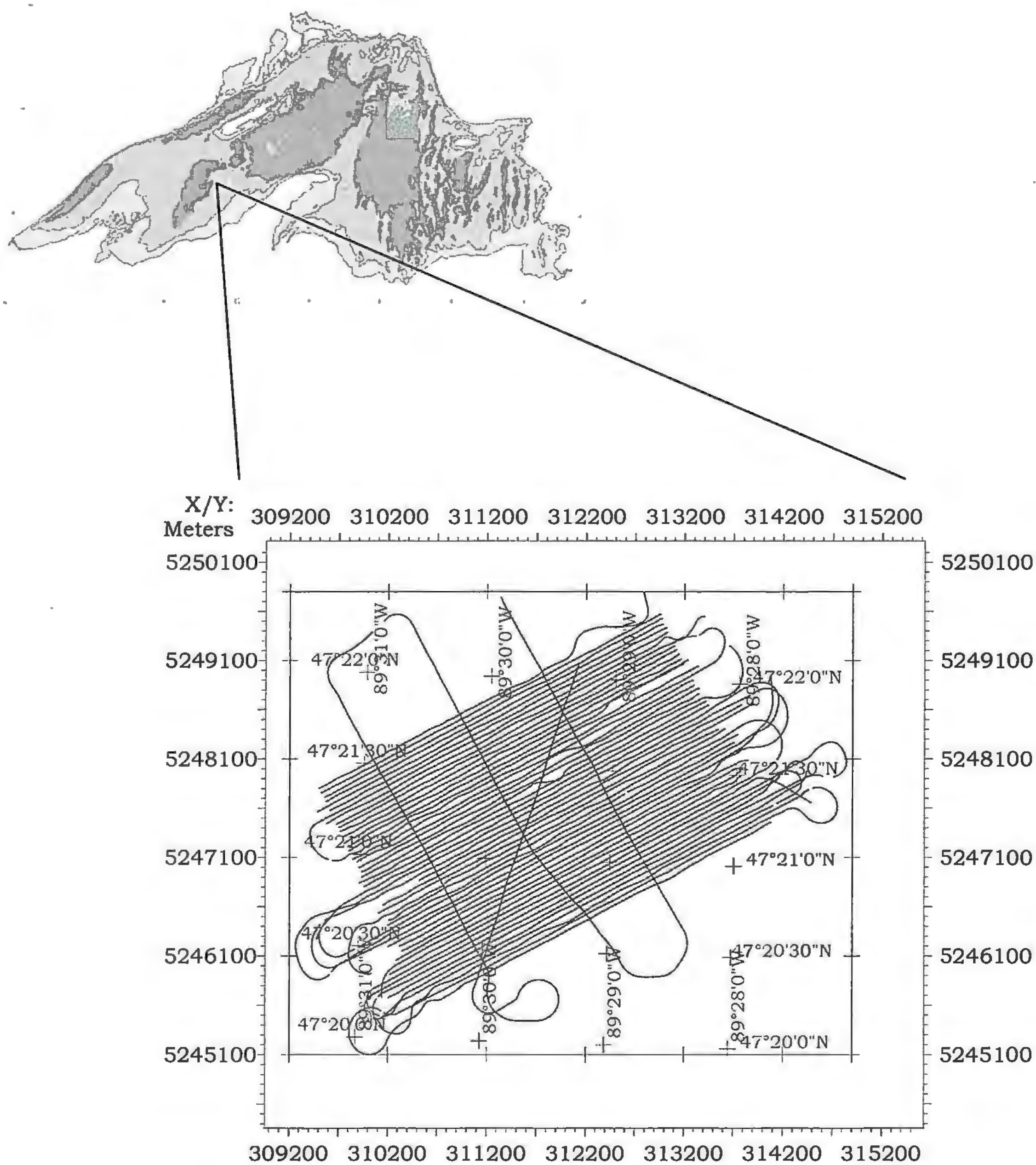


Figure 3.2.1 base map of seismic survey lines

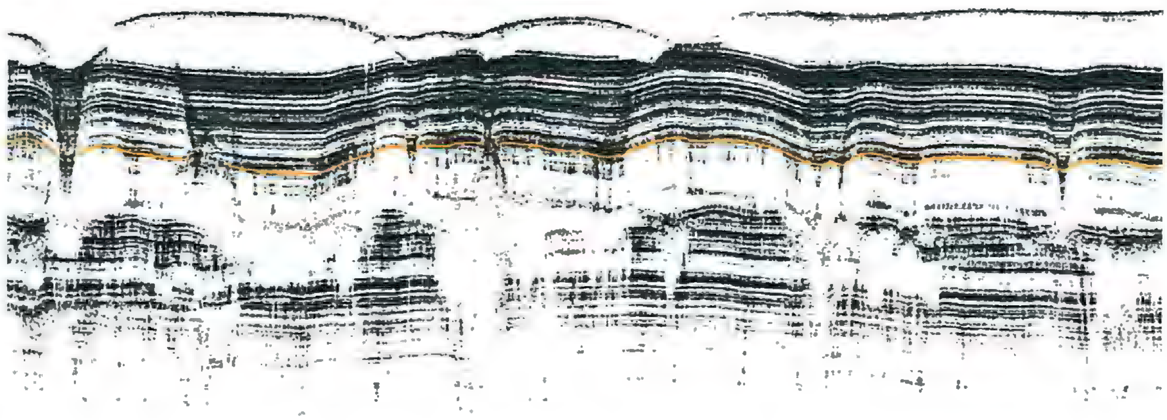


Figure 3.2.2 Line 34 with sample picked horizon

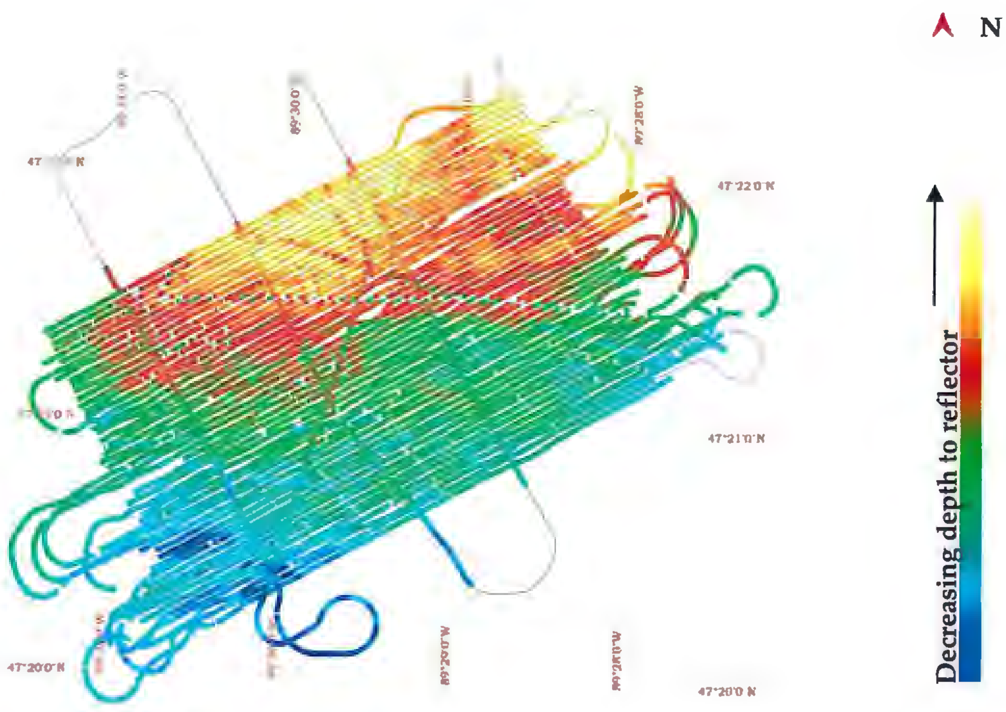


Figure 3.2.3 Orange horizon correlated across study area (pseudo 3-D).

3.2 Sediment Core Recovery and Processing

Three long (6-8m) piston cores were taken from the seismic study area for the purpose of correlating seismic-stratigraphic units with lithological units. Core sites were chosen on the basis of the seismic reflection data in order to sample specific seismic units where they occur near the lake floor.

A modified Kullenberg piston corer was used for sediment recovery. This system is comprised of a steel barrel lined with a 2.75″ polycarbonate liner, a barrel piston, a series of lead weights at its top – all with cable slack, and a triggering mechanism in the form of an arm with a short gravity corer suspended approximately 10 to 12 feet below the bottom of the main core barrel. When the gravity corer contacts the lake floor, weight is taken off the triggering arm, causing cable slack to be released and the piston corer to free fall, driving itself into the sediment. The piston is drawn up under vacuum pressure, pulling the sediment with it, without disturbing the structure of the sediment. The coring apparatus is then recovered by winching it to the surface (Figure 3.2.1).

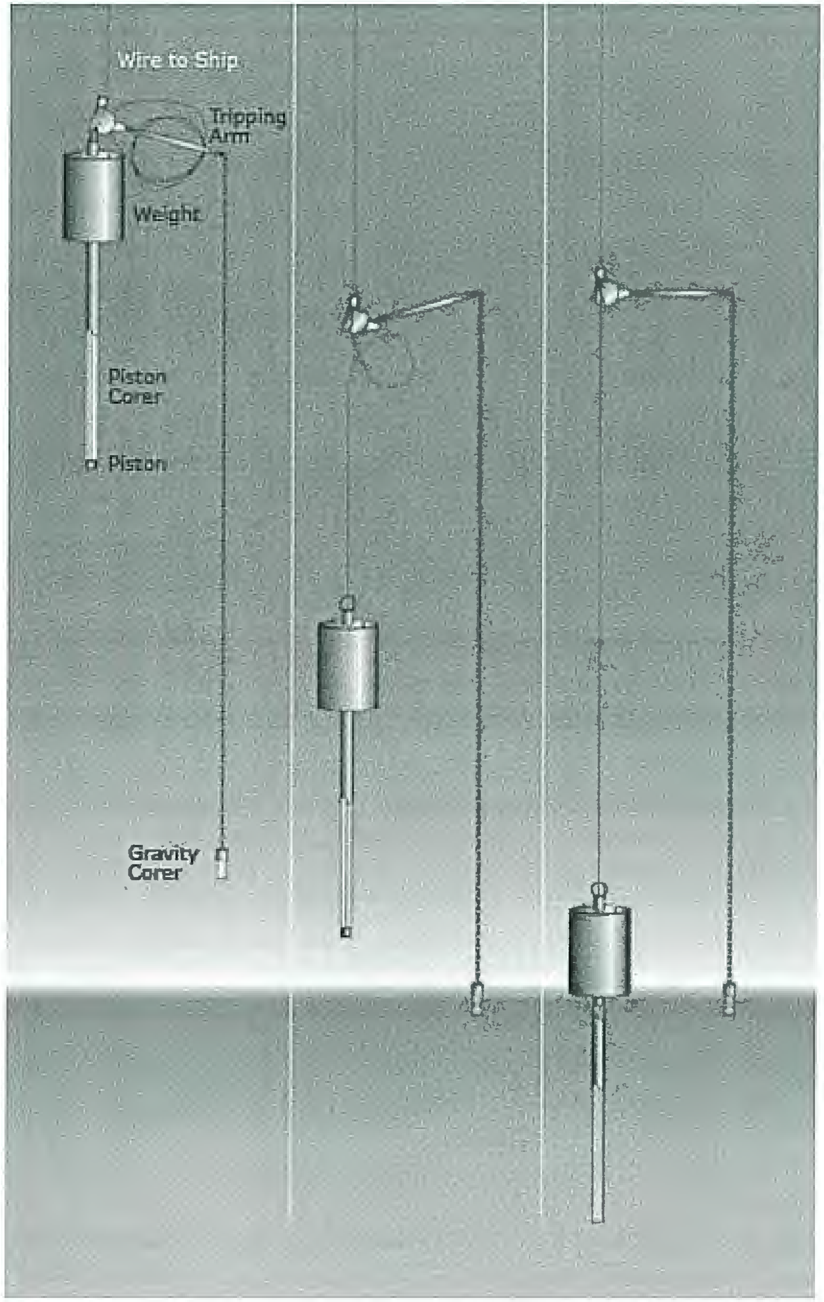


Figure 3.2.4 Kullenberg gravity corer.

The recovered sediment cores were logged for physical properties (including magnetic susceptibility, gamma-ray attenuation for density, and p-wave velocity) on a Geotek core logger at the Limnological Research Center core lab in Minneapolis, MN. Sediment porosity and velocity data can be used to generate synthetic seismic reflection profiles for comparison with actual seismic reflection profiles at each core site. After splitting, the cores were described and photographed so that lithological units can be compared with the seismic reflection data.

Chapter 4: Results

Interpretation of the seismic data follows Stoker, Pheasant, and Josenhans' "Seismic Methods and Interpretation" in *Glaciated Continental Margins An Atlas of Acoustic Images* (1997). Seismic energy is reflected at sediment (or rock) boundaries of differing acoustic impedance. Therefore, seismic reflectors are interpreted as bedding surfaces/sediment horizons.

The character of a seismic reflection can be described in terms of its amplitude, frequency, and continuity. The amplitude, or strength, of a reflection is dependent upon the impedance contrast between strata and is described as low, moderate, or high. Reflection frequency is a function of bed thickness and is described as broad, moderate, or narrow. The continuity of a set of reflections is related to the environment of deposition. A sequence of highly continuous reflections may be indicative of a low energy depositional environment, while heavily disrupted/low continuity reflections are often an indication of a high energy environment. The presence of gas within the sediment can disrupt seismic reflections and cause acoustic blanking.

The configuration of acoustic reflections can be used to infer depositional and post-depositional processes. The three main configurations are: stratified, chaotic, and reflection-free. Parallel to sub-parallel stratified configuration is generally indicative of uniform suspension sedimentation under calm conditions.

Variations on stratified configurations may indicate a change in the rate of deposition, a change in water level, differential erosion, current activity, or tectonic activity. Stratified configurations can be further described as even, uneven, wavy, contorted, hummocky/lenticular, or disrupted. Chaotic reflection configuration can be a result of diamicton dominated sequences, mass flow deposits, or channel fills. The absence of reflections is generally interpreted to indicate a uniform lithology, but may also characterize deposits that have been homogenized by a post-depositional reworking process.

The procedure for seismic interpretation requires several steps:

1) recognize and correlate seismic facies;

2) interpret the seismic facies based on their characteristics;

3) construct a chronology based on the interpretation;

4) evaluate the interpretation-in this case with data from sediment cores.

4.1 Seismic Interpretation

Figure 4.1.1 shows seismic line 34 from the seismic data set. This line is a typical example of the seismic data throughout the study area. I have divided the seismic profile into four distinct seismic facies, based on the characteristics previously mentioned.

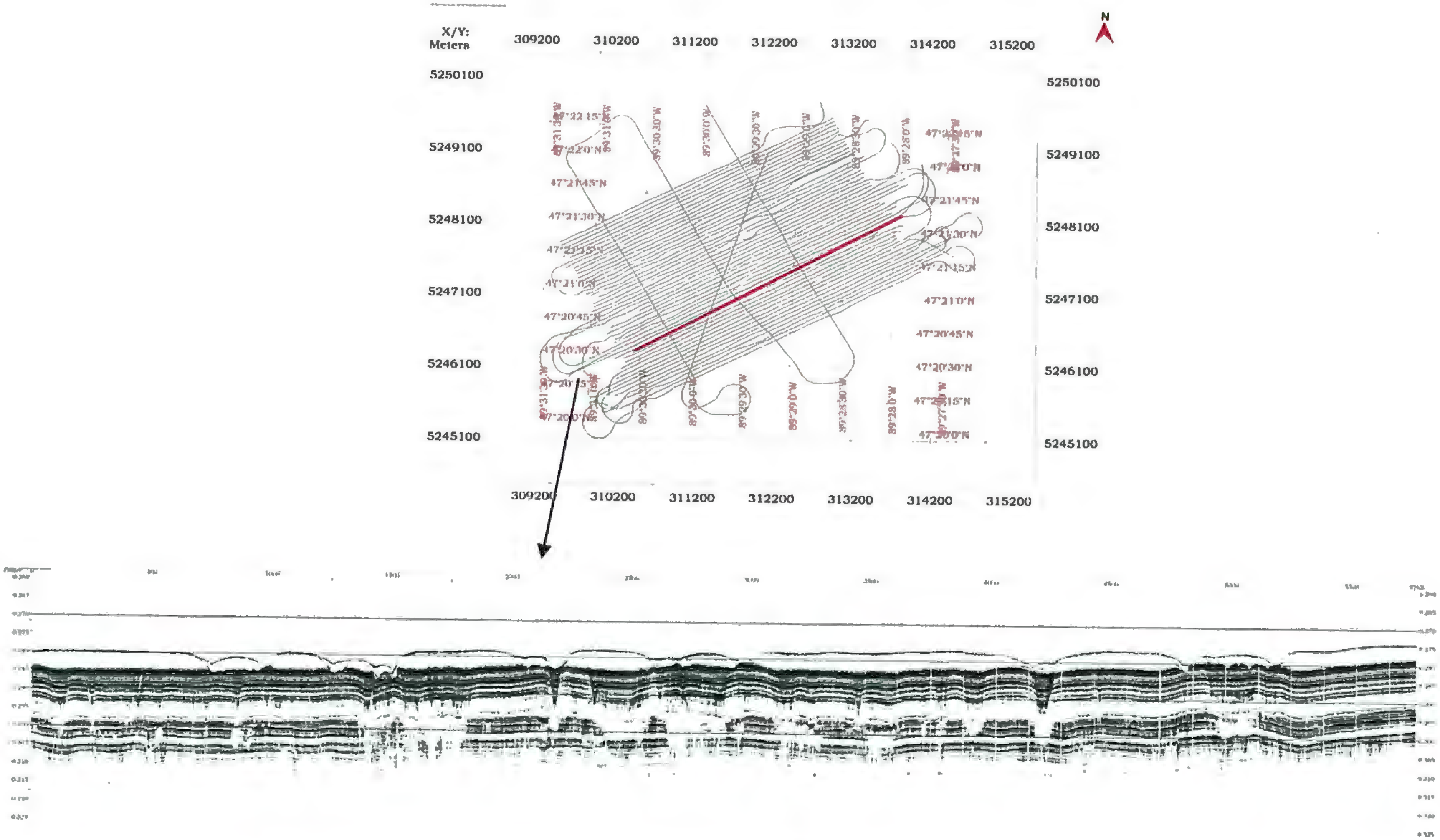


Figure 4.1.1 Seismic data from survey line 34, and location within study area.

Seismic Facies A *(Figure 4.1.2)*

Seismic facies A, the lower most in the seismic section, includes the acoustic basement and is characterized by narrow, parallel to sub parallel, moderate to high amplitude reflections. Reflection strength weakens dramatically below a two-way travel time depth of approximately 0.308 sec (22 – 23 m below the lake floor). Throughout the study area, the acoustic basement averages a time depth of 0.315 sec (28-29 m below the lake floor).

Reflections within unit A are very discontinuous. The unit is characterized by zones of chaotic, low amplitude reflections that are punctuated by regions of spatially limited acoustic blanking. These features are typically much wider than they are high. The scale of the blank zones is highly variable, ranging from a few square meters to more than 100 m across at the widest and up to 10 m deep. They are well developed, especially in the uppermost portions of unit A, near the boundary with the overlying seismic facies B. They are classified as seismic feature A1.

The boundaries of the blanking zones are poorly defined. Reflections on either side of the zone appear to be truncated by the blanking zone, at times appearing to dip down towards its base at its deepest point. The upper surface of the blanking zone is generally conformable with the overlying reflections while the lower boundary is more structurally complex, typically forming a "V" or "U" shape. The spatial distribution of the acoustic blanking regions can be determined by constructing average amplitude maps for time gates defined by mapping horizons in Unit A (Figure 4.1.3). By adopting this strategy it is possible to remove structural complications in the calculation of the amplitude maps. The maps clearly demonstrate that the blanking features, recognizable by their low amplitude in the maps, may be traced across adjacent survey lines, in many cases spanning the entire survey area. Individual blanking features exhibit linear or curvi-linear paths across the survey, frequently intersecting others at high angles.

Careful examination of the reflectors near the top of, and truncated by, the blanking zones reveals that they frequently exhibit lateral thickness variability, which are defined as seismic feature A2. The best example of this behavior is seen in the horizon nearest the top of facies A. This reflector thickens and thins by as much as 50% (Figure 4.1.2). The variation occurs almost exclusively in close proximity to, and on either side of a blank zone feature. These changes in thickness also show a strong correlation with the location of the faulting in the overlying seismic facies B and C.

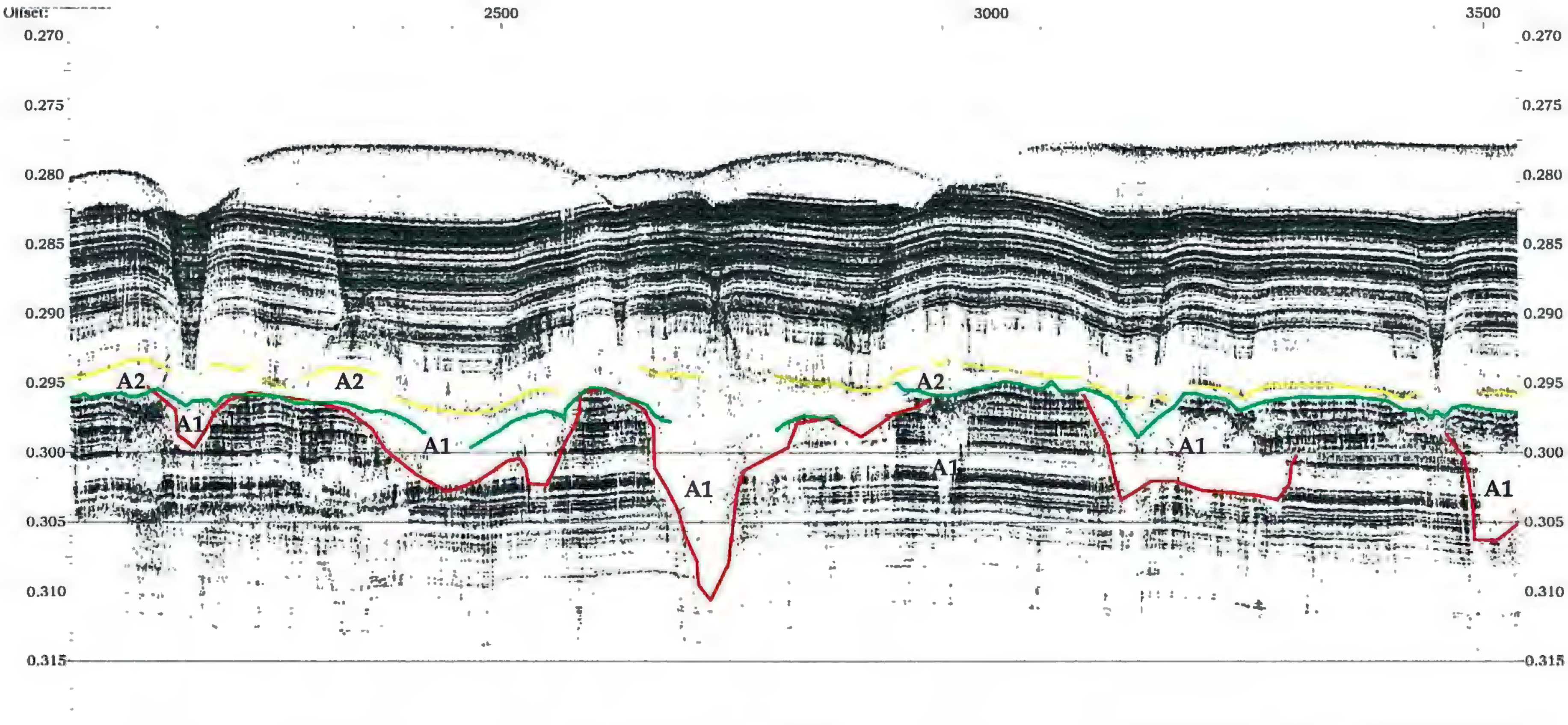


Figure 4.1.2 – Seismic Facies A. Upper boundary marked by yellow horizon, lower boundary is acoustic basement. The lower boundary of feature A1, the blanking zones are marked by the red lines. The yellow and green horizons mark the boundaries of the bed that demonstrates thickness variability, feature A2.

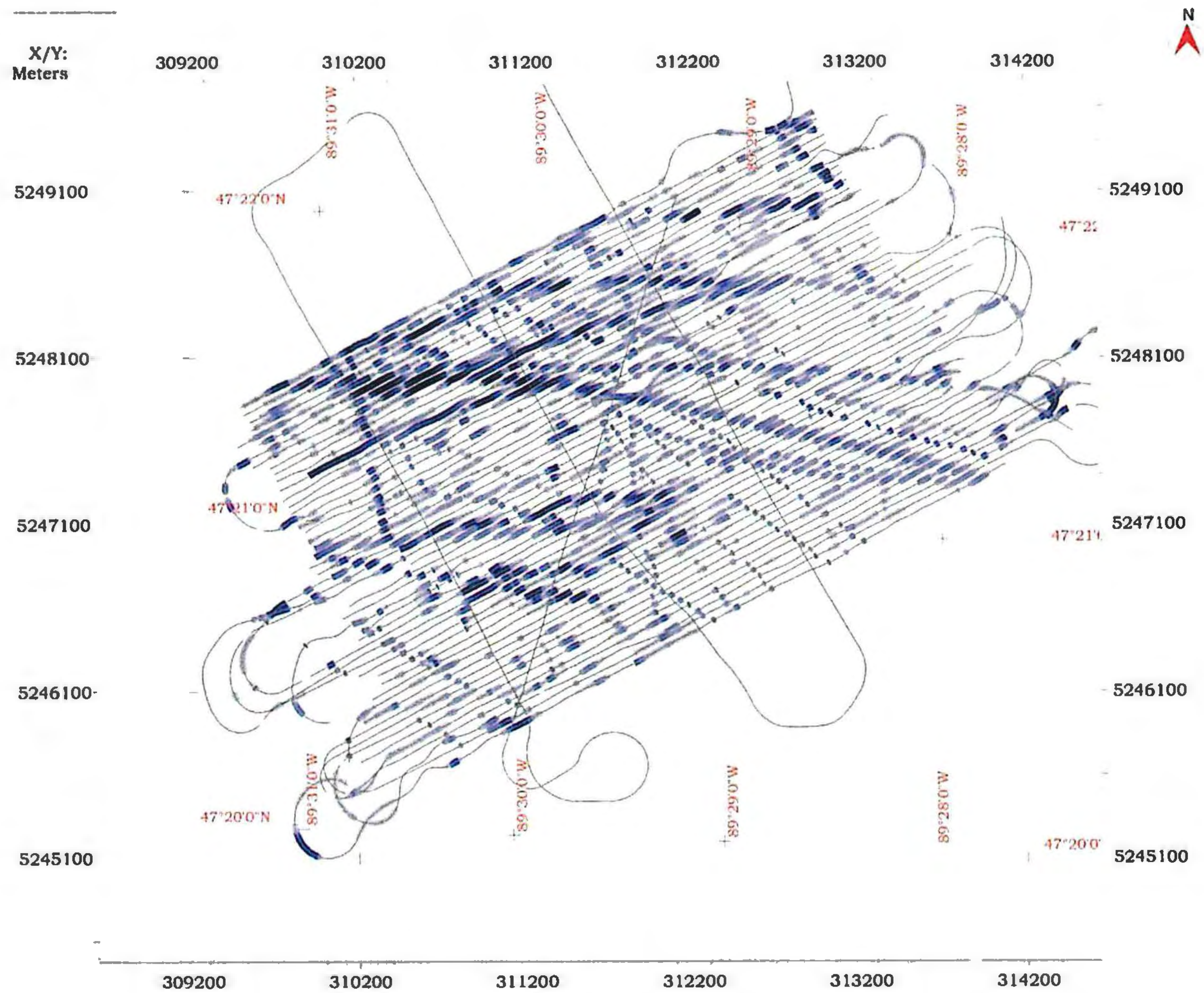


Figure 4.1.3 – Spatial character of the blanking zones in map view. Linear to curvilinear paths are deeper as color becomes darker.

Facies B *(Figure 4.1.4)*

Seismic facies B rests unconformably on seismic facies A; although over significant portions of the survey area it can appear to be conformable. It is characterized by a weak to reflection-free amplitude. Those reflections that are identifiable are generally high frequency and exhibit wavy parallel or weakly contorted character. In some instances, these reflections can be chaotic in nature. Reflection continuity is low in facies B, its internal reflections are often disrupted, especially over the acoustically blank zones in facies A and the zones of faulting and/or fracturing that are more commonly developed in the overlying facies C.

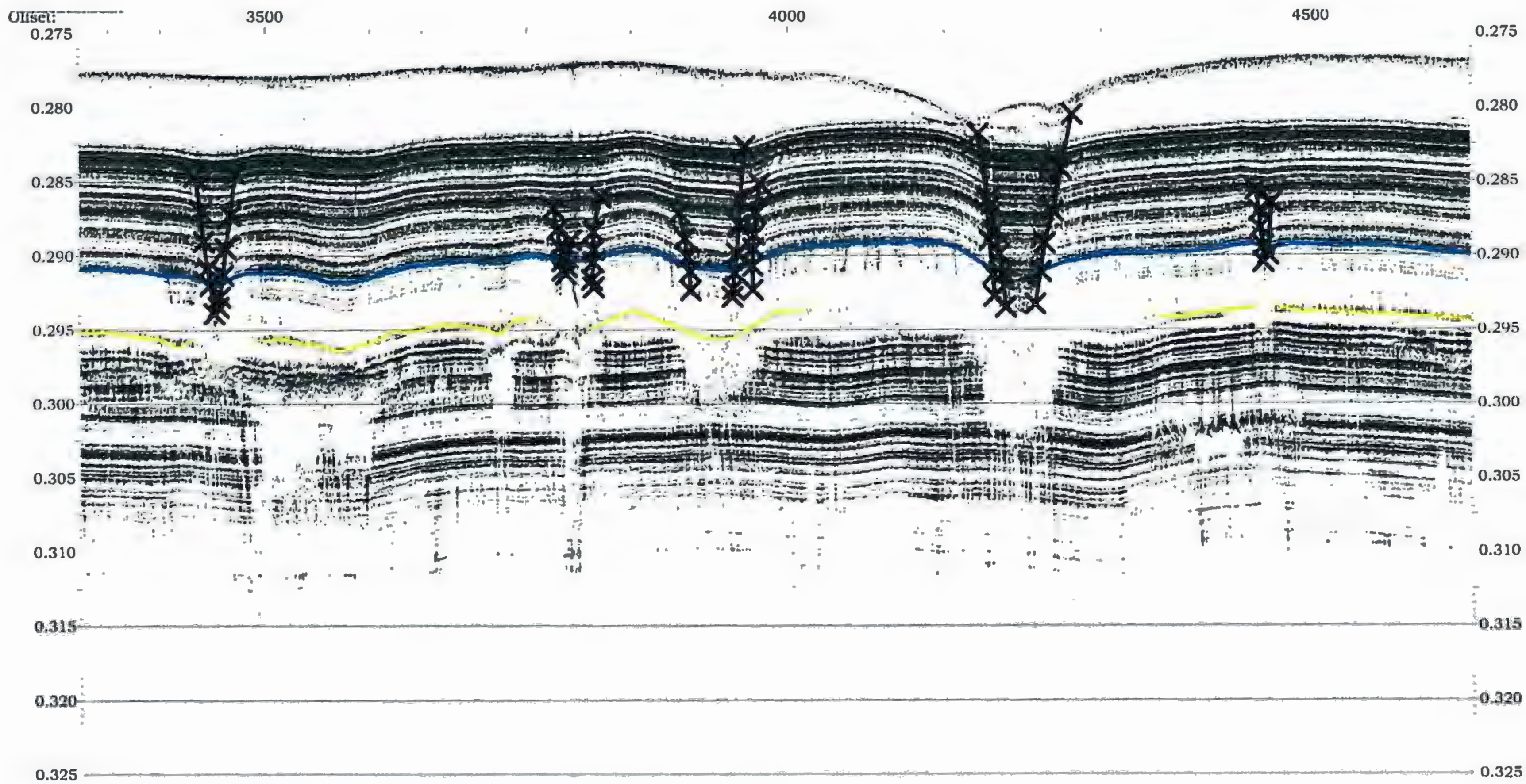


Figure 4.1.4 Seismic facies B – between blue and gold traces. Faulting from overlying facies C shown disrupting weak reflections

Facies C *(Figure-4.1.5)*

Seismic facies C comprises the majority of the upper half of the section, resting conformably on top of facies B. Reflection amplitudes within this section are generally strong. The geometry of the internal reflectors in this facies is parallel to contorted and convergent. These characteristics are frequently disrupted by seismic feature C1. Feature C1 is characterized by the offset of layers by faulting. In map view, these faults take on a curvilinear pattern that appears to correlate with the acoustically blank zones (seismic feature A1) within facies A (Figure 4.1.6). The faulting is typically normal with apparent throws of less than 0.5m. Three styles of faulting (Figure 4.1.7) are recognized: monclinal, graben-like and conjugate pairs.

The interpretation of these structures is complicated by the seismic data's limited ability to resolve small scale features and the presence of imaging artifacts associated with the low relief flexure of the sediments. The contorted/convergent character of the reflectors in facies C is often associated with the faulting in the section. In many cases, the flexure appears to increase in severity with increasing depth and continues into the underlying facies B. The increase in flexure severity may indicate that the faults are growth faults, that remained active throughout the period of deposition as opposed to forming by an abrupt event sometime after the full sequence of sediments were deposited.

It should be noted that there is a significant amount of vertical exaggeration in the included figures. The aspect ratio of the horizontal to vertical scale is more than 16:1. This makes the angular relationships more obvious for interpretation, but overemphasizes the actual low relief of the structures.

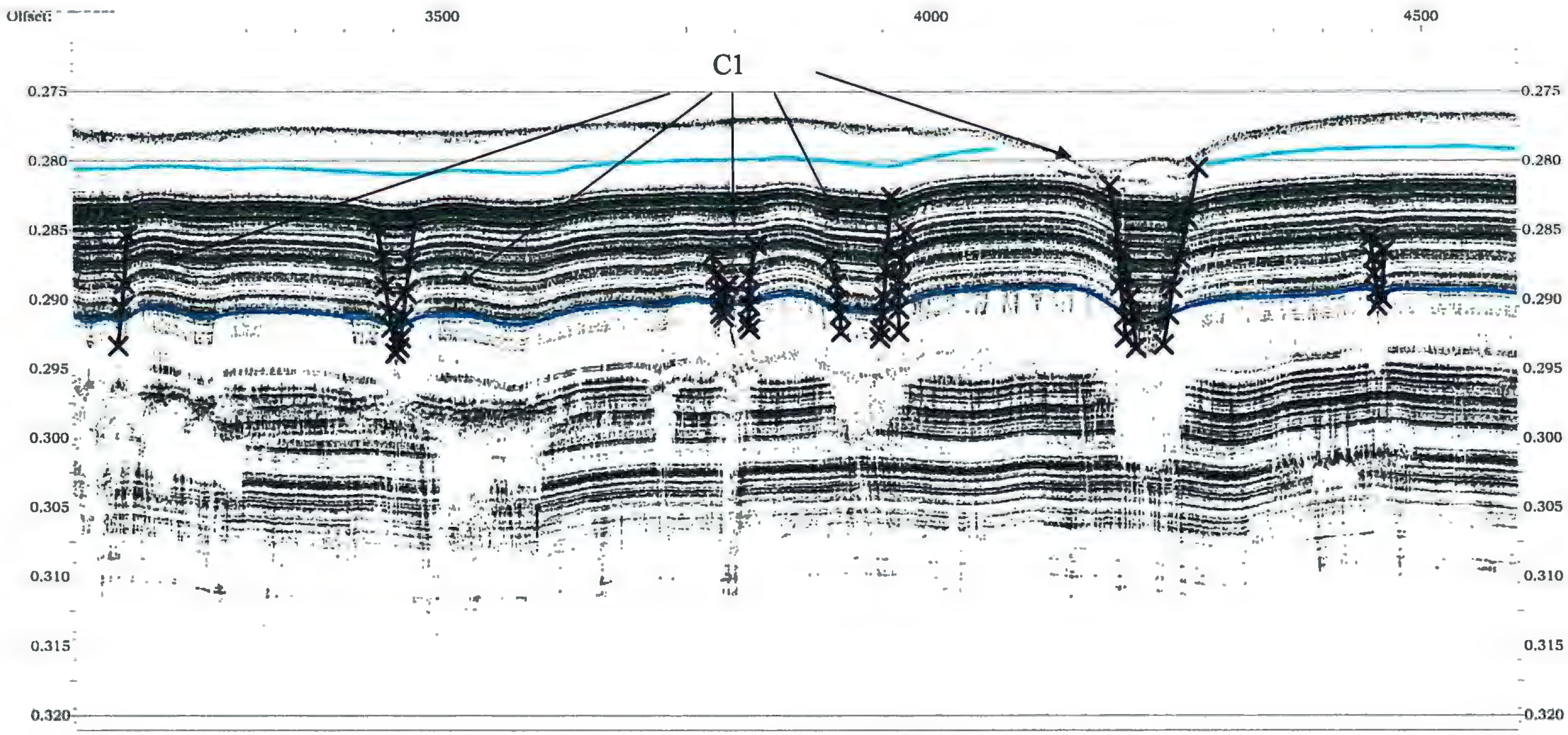


Figure 4.1.5 – Seismic facies C with feature C1, normal faulting, labeled.

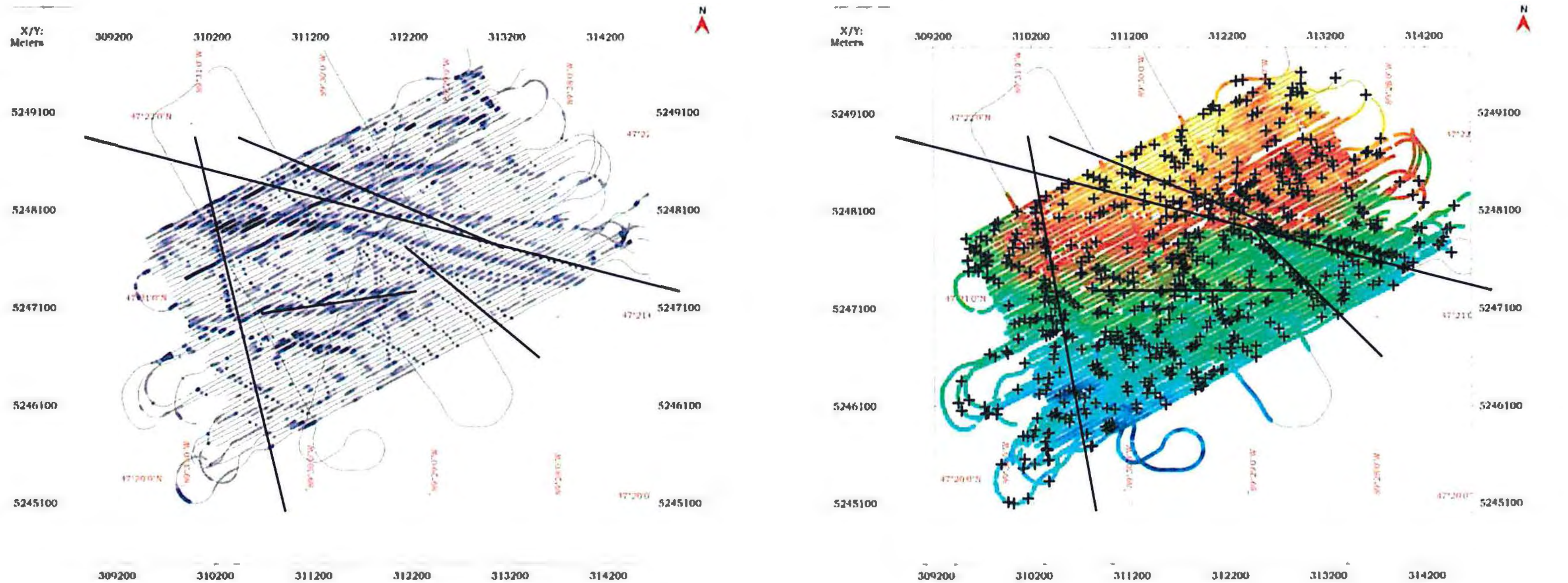


Figure 4.1.6 – map view of linear character of acoustic blanking in facies A (left) and fault traces in facies C (right). Lineations of correlative nature have been highlighted with trace overlays.

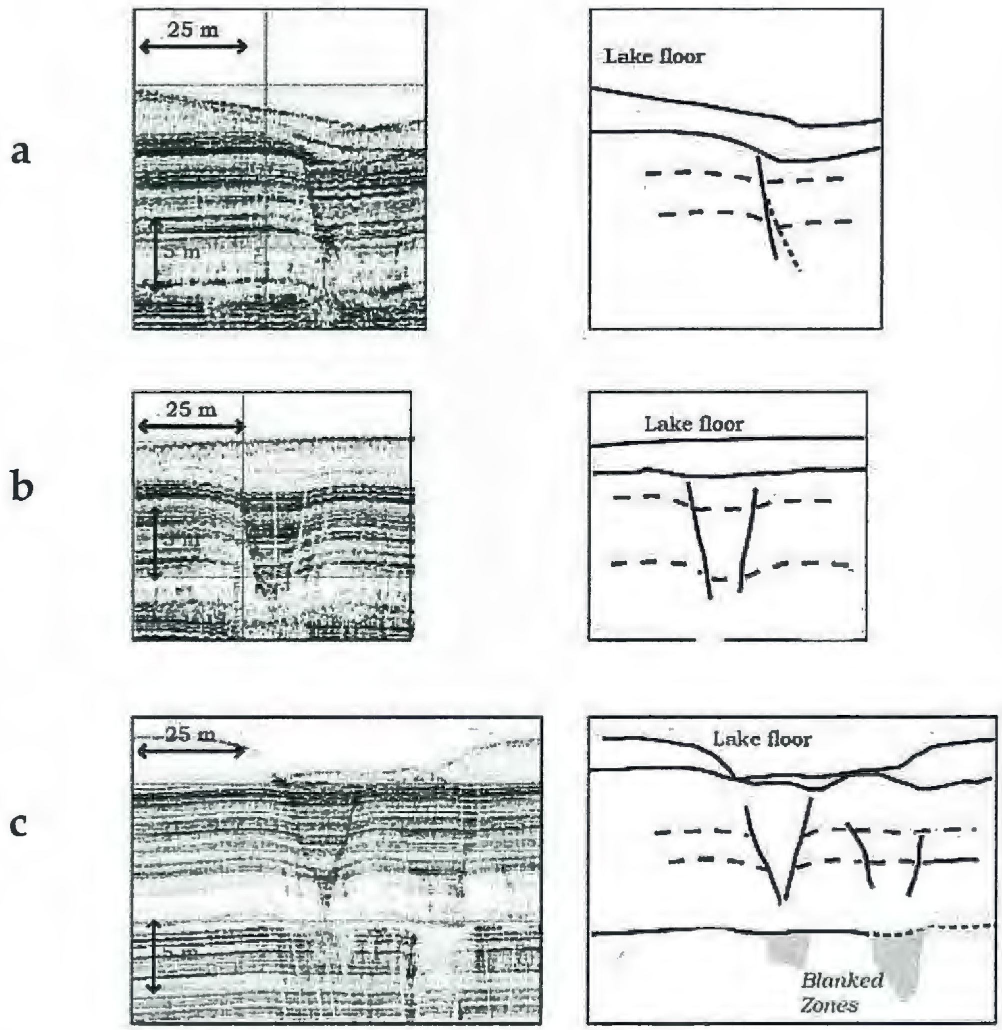


Figure 4.1.7 – Examples of the three styles of faulting recognized in facies C. Note the flexure of the reflectors adjacent to the faults; it appears to increase with depth along the fault. The acoustic blanking is also apparent and correlative to the faults. (a) monoclinal, (b) graben-type, (c) conjugate pair.

Facies D *(Figure-4.1.8)*

The uppermost seismic facies, facies D, is characterized by a lack of internal reflections and pronounced discontinuity of the entire facies. The upper boundary of facies D is the lake floor. The discontinuity is caused by distinct thickness variation within the facies. Correlated across the study area and shown in map view, this thickness variation expresses the ring pattern, previously mentioned in chapters 1 and 2.5, on the lake floor. The rings tend to be roughly 100m in diameter, with troughs 5-10m wide and 5-10m deep (Figure 4.1.9). In map view, both the bounding reflector between facies D and C and the upper boundary of facies D (the lake floor) exhibit the ring features. In seismic profile, there is a correlation between the thickness variation of facies D and the faulting within facies C.

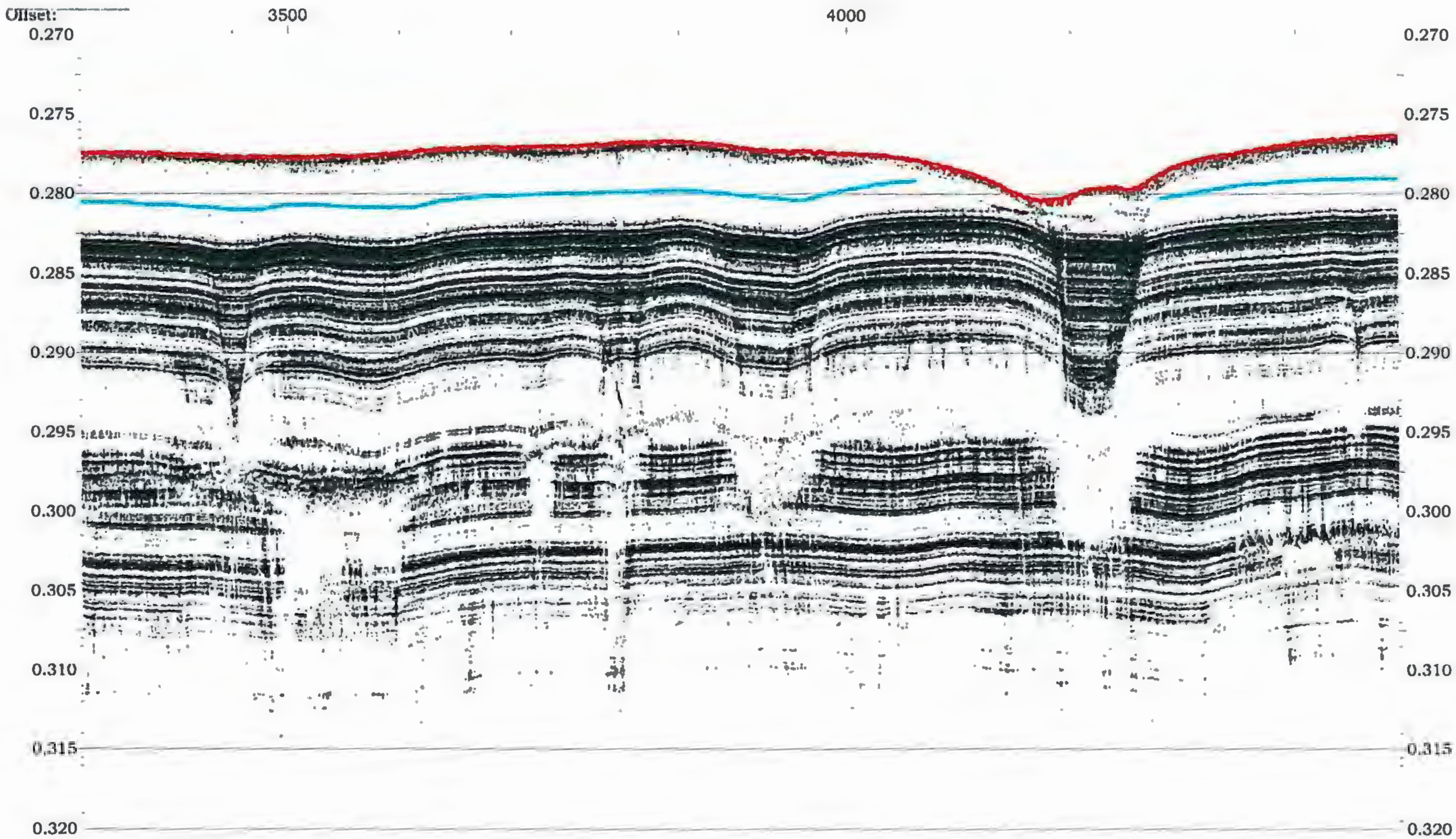
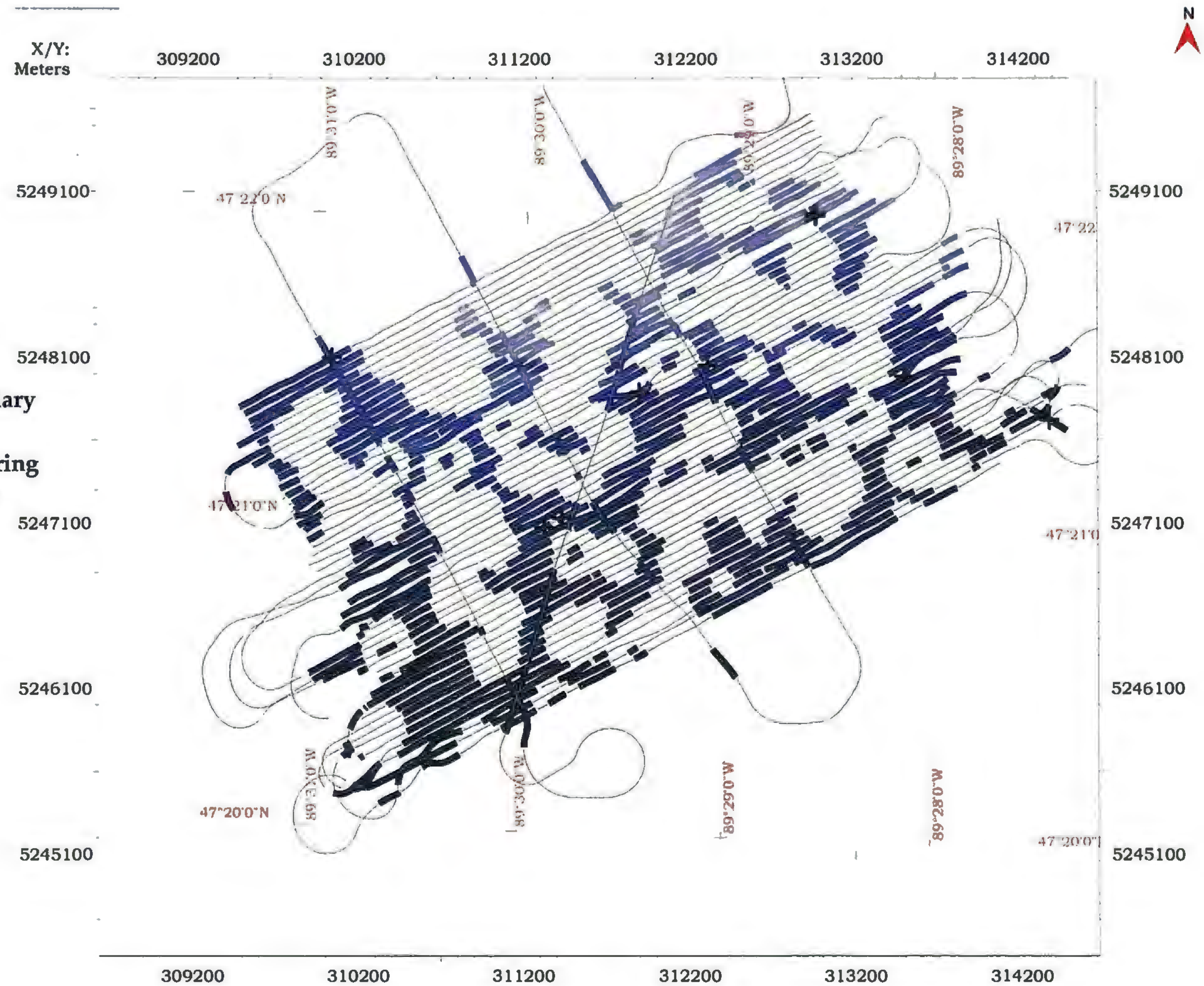


Figure 4.1.8 – Facies D, with upper boundary in red (lake floor), and lower boundary light blue.

Figure 4.1.9 – spatial character of lower boundary of seismic facies D. Truncations correlate to ring pattern on the lake floor.



4.2 Sediment Core Interpretation

Seismic interpretations were validated by taking sediment samples at appropriate locations within the study area. In fine grained glacio-lacustrine clay and silt, gravity coring is a very effective sediment recovery method. Core BH02-1P, described below, yielded a well intact record of sediment. Cores 2P and 3P did not recover any sediment below the grey varves.

4.2.1 Core BH02-1P

911 cm below the lake floor, sediment consists of pink to red very fine sand and silt to silty clay. It is varved and has bed thickness up to 3.5 cm. These varves begin to thin 69 cm from the bottom of the core to a thickness of ~ 1 cm. 825 cm below the lake floor, the sediment color begins to change to red-grey, with varve thicknesses of 0.5 cm. The consistency of the varved sequence is interrupted at a depth of 793 cm by poorly stratified to massive red silty clay. This red sediment transitions to a thickly stratified varve sequence with beds 4cm thick, again thinning to 0.5 cm and remaining consistently stratified for the next 140 cm.

At a core depth of approximately 653 cm, the red-grey silty clay becomes predominantly grey clays, with interbeds of red over 40-45 cm. The next 40 cm are jumbled grey-red varves with a massive wedge of red clay with blebs of grey clay (Figure 4.2.1). The grey sediment is generally finer grain size-silty clay to clay.

Above the massive red clay wedge, the sediment abruptly becomes predominantly grey varved clay with bed thickness of .25 to 1 cm. The varves are consistent in thickness and continue to the top of the section. Remnants of red clay can be seen in the darker portion of the varves, but the concentration decreases to zero 427 cm below the lake floor. The grey varves continue through to the top of the core. Farrand (1969) suggests that the change in sediment color is a representation of glacial retreat from the basin. The grey sediments represent material derived from north of the basin that would not have been available for deposition until the ice had retreated far enough to expose them. Low angle normal faulting appears about 400 cm below the lake floor, and becomes increasingly extensive to the top of the core (Figure 4.2.2). Fault throws are on the order of 0.5 cm.

Figure 4.2.1 Massive red clay wedge near 570 cm depth in BH02-1P, and jumbled bedding below.

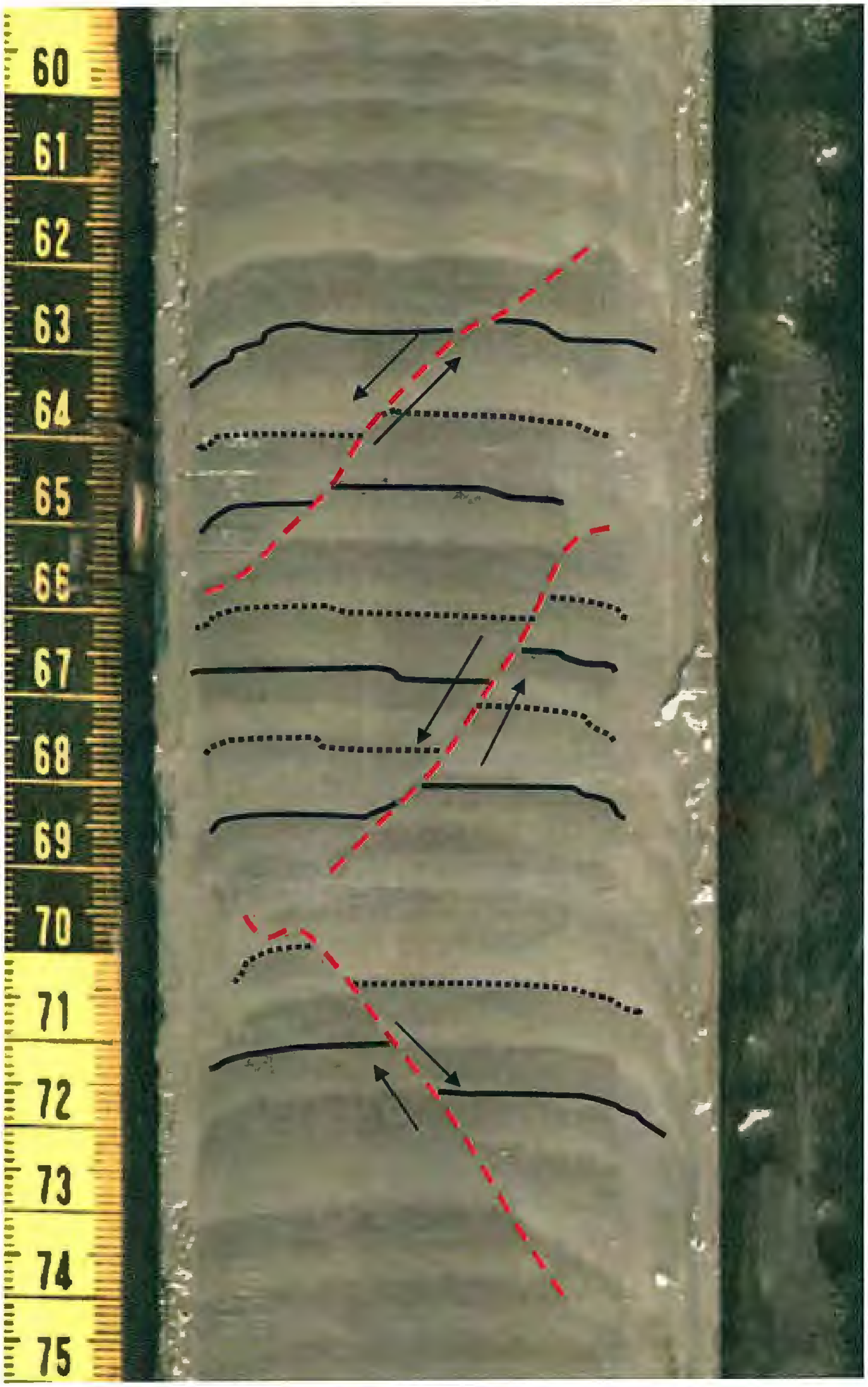
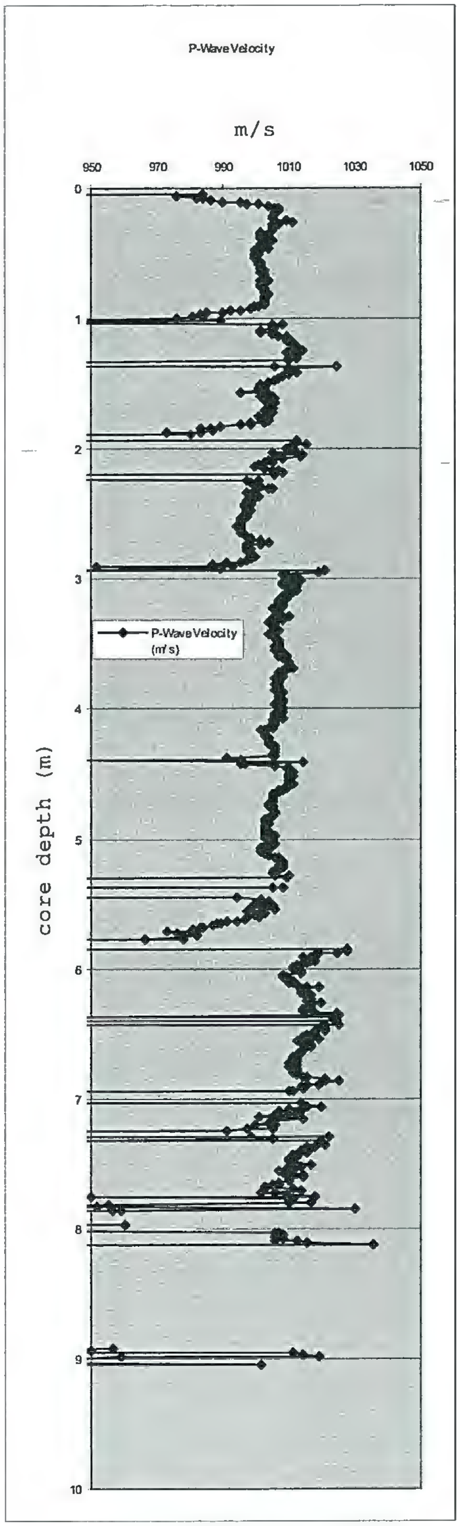


Figure 4.2.2 Core scale faulting near top of BH02-1P.

4.2.2 Geotek Data

P-wave velocity among the sediments decreases as the predominance of grey sediment increases (up in the core). The density of the sediments also decreases with the transition from red to grey. A comparison between the visual description of the core and the density data (Figure 4.2.5) shows that the highest average density correlates to the uppermost portion of the red varved sediments.

Acoustic impedance is the product of density and velocity, so changes in either of these factors would be expected to generate surfaces that are capable of producing seismic reflections. Since the lithology of the sediments remains consistent throughout the core, it is assumed that the subtle changes in velocity and density are the result of the slight change in grain size and variations in the porosity/pore water content of the sediment.



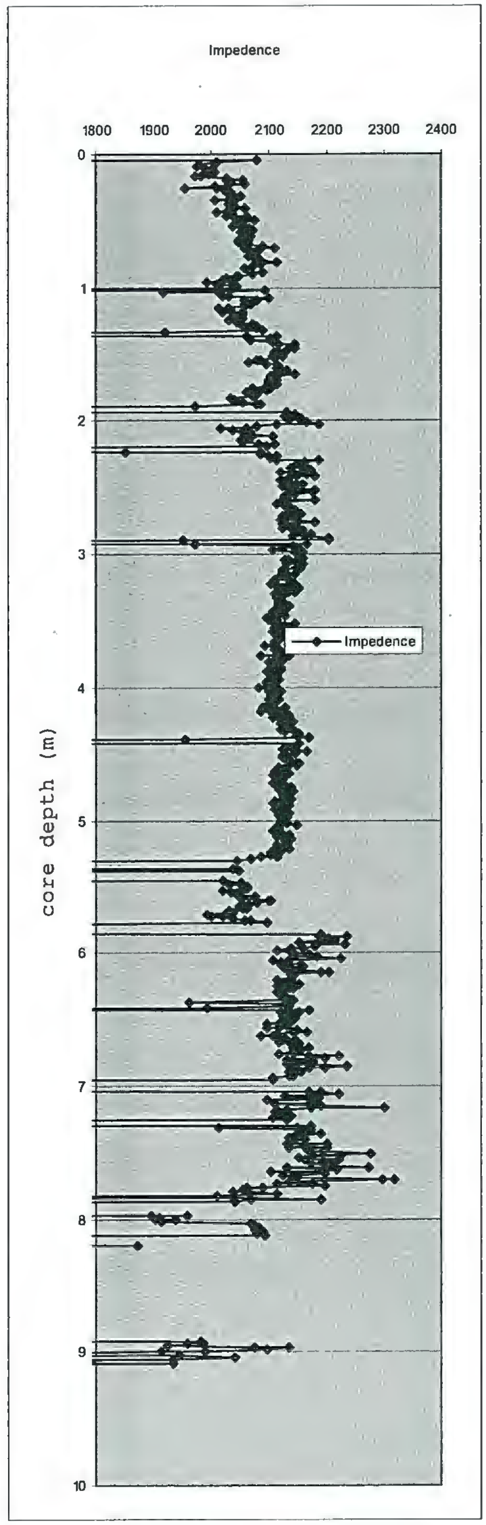


Figure 4.2.3 Charts of Geotek data for P-wave Velocity and Acoustic Impedance.

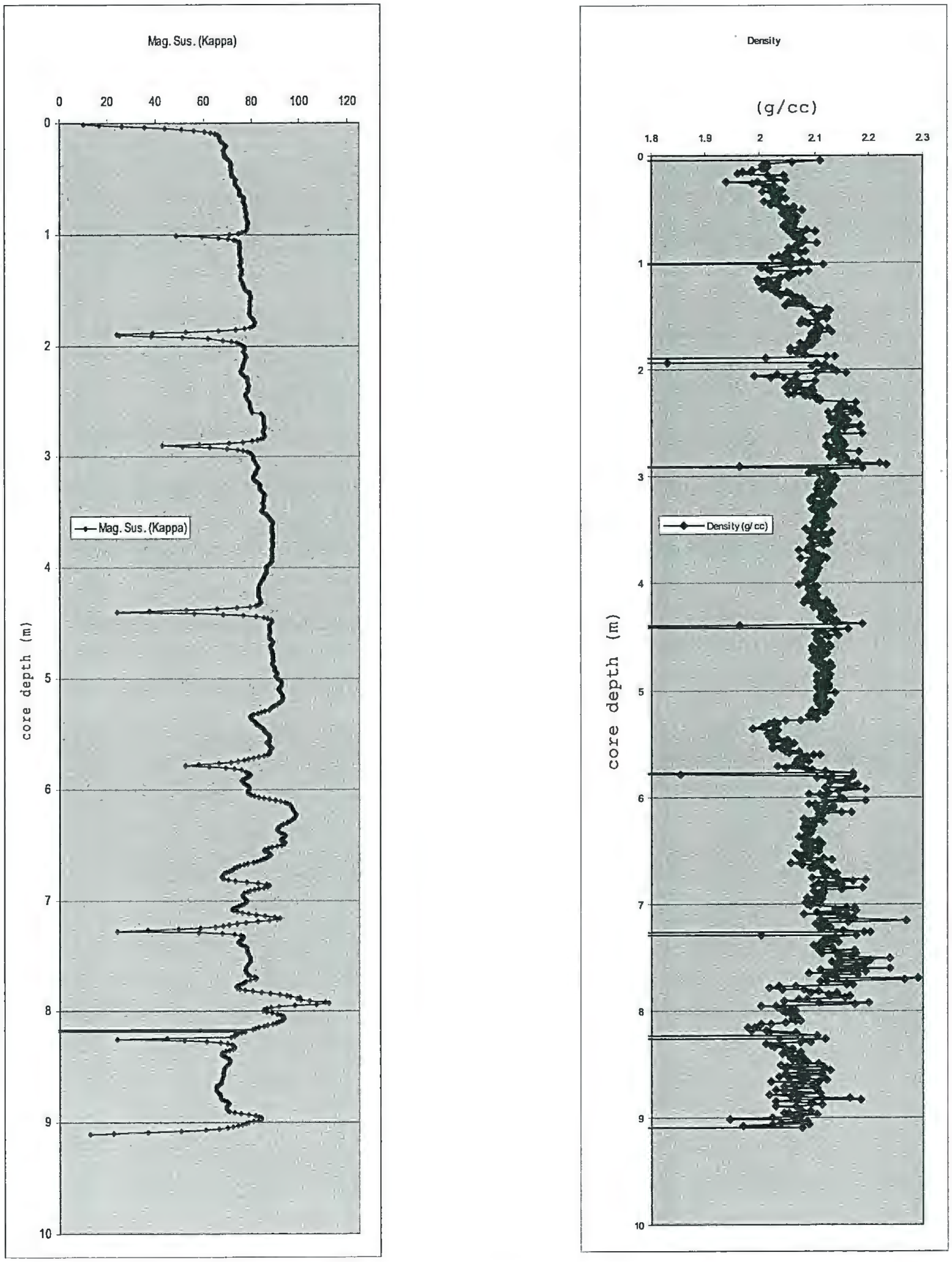


Figure 4.2.4 Charts of Geotek data for Magnetic Susceptibility and Sediment Density.

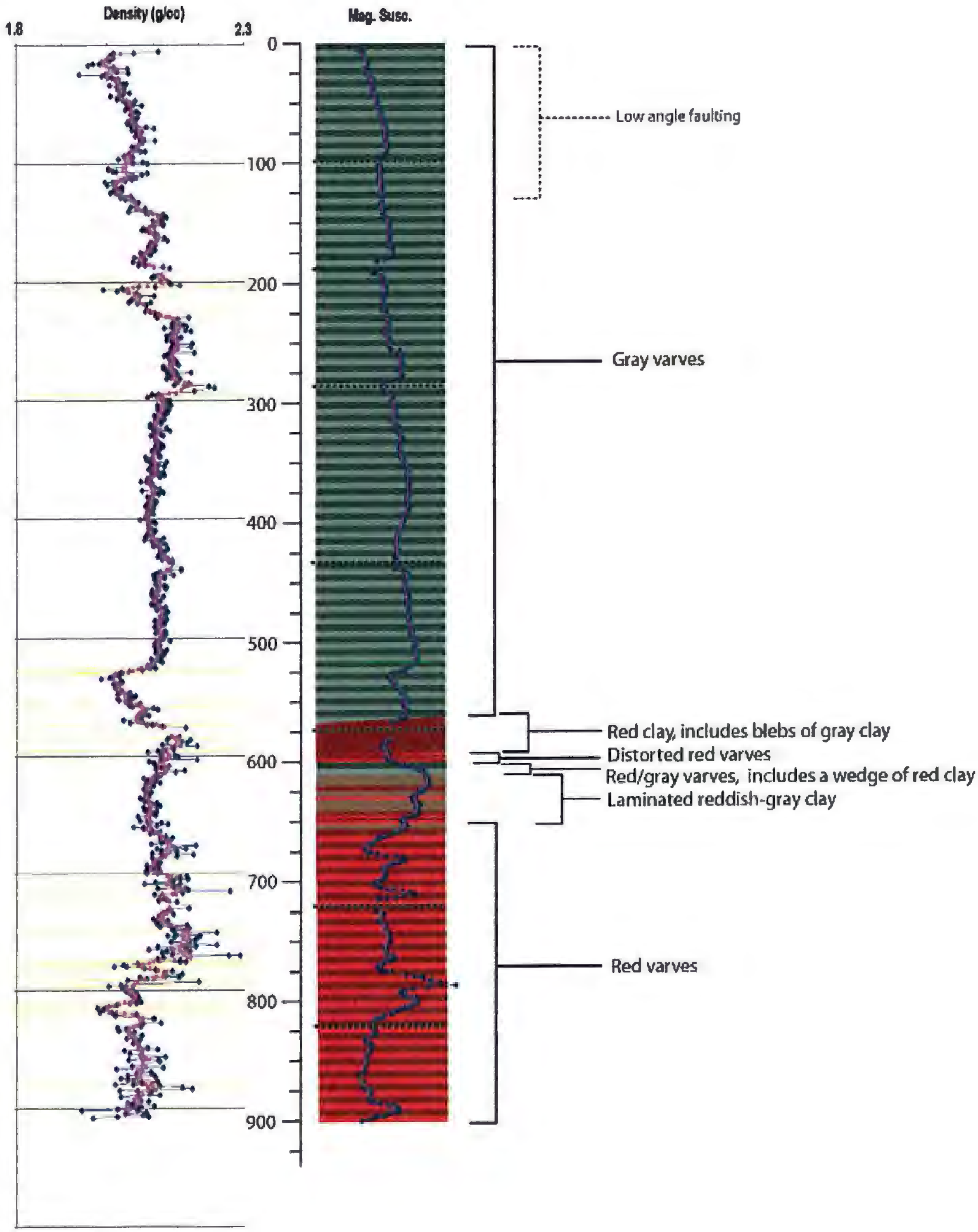


Figure 4.2.5 Comparison between visual core description and Geotek data for density and magnetic susceptibility.

Chapter 5: Discussion

In Chapter 1 I posed a number of questions regarding the nature of the recent sedimentary record in western Lake Superior. In this chapter, I return to these questions, using my observations to address each in turn.

-Has post-depositional remobilization of the glacial sediments occurred?

The depositional environment of glacio-lacustrine silts and clays in the Lake Superior basin is typically calm, deep water – leading to originally horizontal bedding planes. There is distinct evidence that these sediments have experienced post-depositional disturbance and remobilization. Examples to this effect can be seen in both the seismic data and cores.

Seismic evidence of remobilization is seen in several features. These include: ring-shaped land forms of the lake floor; apparent normal faulting of the reflectors within facies C; lateral thickness variations between reflectors; V or U shaped "blank zones" – areas of little to no internal reflection within facies A; an unconformable boundary between facies A and B; and discontinuous, wavy, contorted and jumbled or chaotic reflections within facies A, B, and C. The fact that there is no preferential deposition in the low spots of contorted reflectors indicates that the deformation took place sometime after original deposition. This is further evidenced as the flexure of reflectors tends to decrease upward in the seismic section.

Within the core samples, evidence of post-depositional disturbance of the sediments includes offset of the beds with core-scale faulting within the grey varves and wedges of unstratified red silt-clay appearing near the transition from red to grey sediments.

By correlating the seismic and core data (Figure 5.1), it is apparent that many of the features attributed to remobilization occur within the red sediment. Since the cores failed to penetrate to a depth that fully samples the red sediments, many of the features observed in the seismic section were not recovered.

The seismic data clearly indicates that these individual features are spatially connected. Blanking zones are often flanked by beds that show lateral variations in thickness, thinning or even disappearing at the blank zone and swelling to greater than normal thickness either side of the zone. Reflectors intersecting the blanking zones frequently exhibit a wavy, contorted character that increases in intensity close to the blanked zone. Maps of the fault intersection with the gray varves show strong visual correlations with the blank zones.

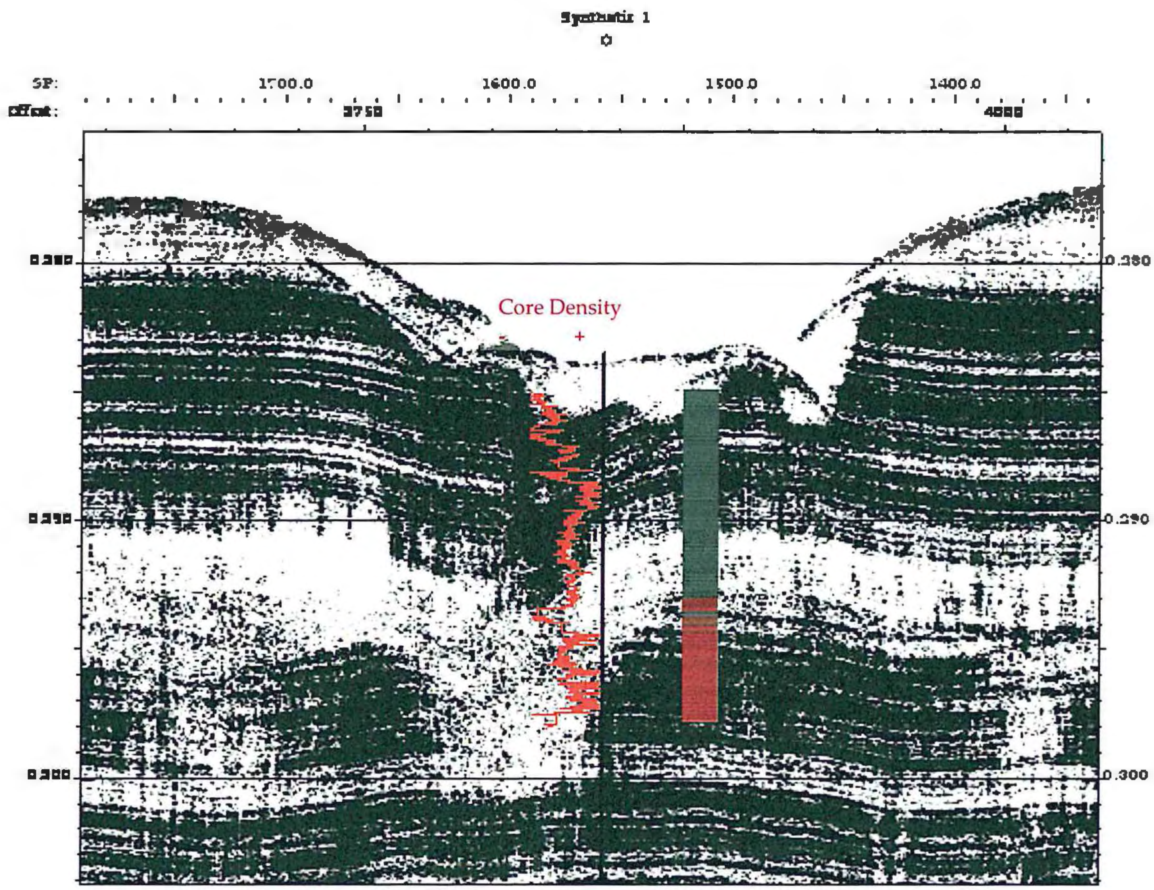


Figure 5.1 Synthetic core log and density profile correlated with seismic profile

-How do these remobilization events fit into the established glacial history of the Lake Superior basin?

The correlated seismic and core data show that the majority of sediment remobilization has occurred within the upper portion of the red varved sediments. Farrand (1969) has attributed the transition from red to grey sediments as the record of ice leaving the Lake Superior basin. The red sediments indicate the erosion, transportation and deposition of material within the basin itself. The grey sediments are representative of lithologies from outside the basin. Since ice was still in contact with the basin during the period that the red varves were deposited, glacial processes probably played a significant role in the remobilization of the sediment.

While ice was in contact with the basin, the drainage of Glacial Lake Agassiz underwent major changes. As the LIS retreated from the area, the eastern GLA outlet through the Great Lakes opened and closed with ice retreat and advance, causing several major Lake Superior level fluctuations. These outbursts had the potential to dramatically increase sedimentation rates, sediment pore water content, and cause episodes of increased iceberg calving from the Superior Lobe margin. Breckenridge (2005) and Teller et al. (2002) argue that the majority of eastern GLA outbursts through the Great Lakes occurred between 10.6 ka and 9.2 ka, when the Superior Lobe left the basin for the final time.

-What is the mechanism for the post-depositional remobilization?

Given the ice-dammed glaciolacustrine setting, the most plausible interpretation for the origin of the blank zones in the seismic section is iceberg scour. The linear nature of the blank zones along with the established lack of trapped gas within Superior sediments discounts the possibility that the acoustic blanking is the result of gas. More likely, the lack of internal reflection is the result of mechanical homogenization of the sediments. When sediment is homogenized, any prior differences in density and velocity are reduced or destroyed, erasing any change in acoustic impedance, resulting in no reflective character (Stoker et al. 1997).

The variability of bed thickness and flexure near blank zones further supports the theory of iceberg scour. A central characteristic of iceberg scour is the formation of lateral and frontal berms of sediment. This is the result of the keel of the iceberg plowing into the lake floor sediments, pushing at the front and piling and displacing sediment at its margins (Figure 5.2). Eden and Eyles (2001, 2005) demonstrate that iceberg scour can affect sub-scour sediment to a depth of 2 to 4 times the scour depth in the Sunnybrook deposit in southern Ontario.

Figure 5.2. Schematic sketch illustrating the sequence of events leading to the formation of the scour-related features. a). Iceberg plows through soft lake floor sediments, pushing them aside and building lateral berms, b). Rapid deposition of additional sediment blankets the scour, quickly erasing it's expression, c). Subsequent compaction of the ice-turbated fill in the scour triggers collapse of the overlying sediments that is frequently associated with paired groups of inward directed faults.

The wavy and contorted nature of the overlying beds is the result of deposition on top of the scour berms and above the blank zones of scour (Figure 5.3). In the sediment of Glacial Lake Rintel in NW Germany, Winsemann et al. (2003) observed normal faulting within iceberg scour fill, and in overlying and adjacent deposits. They also noted the presence of small grabens and "pronounced downwarping of the overlying strata into the central depression." This fact further ties the extensive normal faulting observed in our study area to the underlying blank zones.

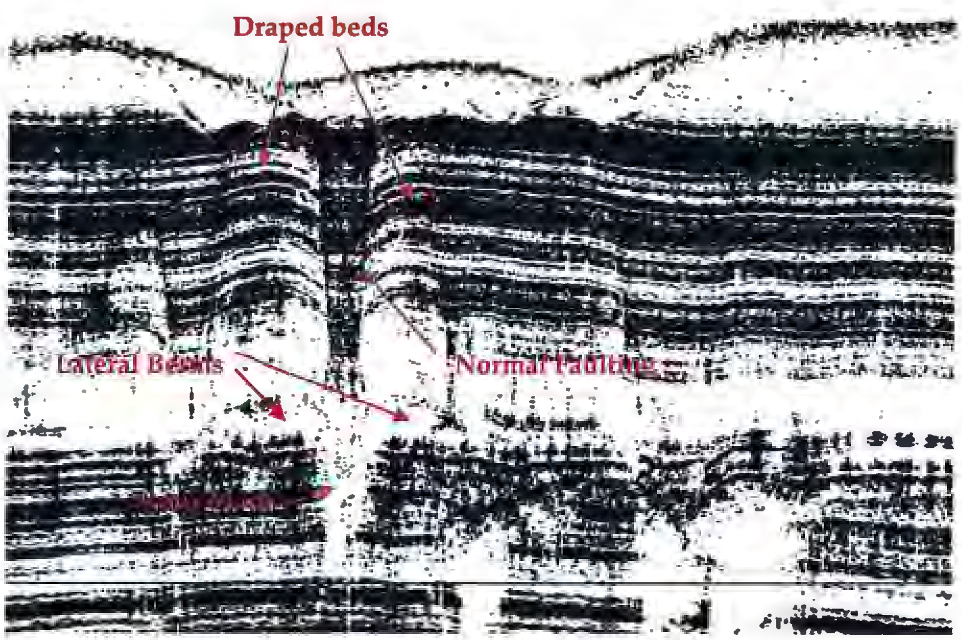


Figure 5.3 Overlying beds draped over scour berm.

In the Winsemann study, the collapse of sediments in the scour trough was interpreted as the result of melting dead ice anchored in the sediment. As the trapped ice melted, space was opened for overlying sediment to fall into, causing the downwarping and extensional faulting. This would indicate the faulting should be localized over the terminus of the scour mark. In the case of the Lake Superior sediments, faulting is linear and often follows the full trace of the scour, indicating that dead ice did was not the culprit for sediment collapse. It is more likely that the reworking of the sediments by the iceberg, which created an unconsolidated "fill" in the scour. With the continued supply of sediment rich water draining from the LIS, the scour and the sediment filling them were quickly buried by the rapidly accumulating red varved clays. This blanket of sediment increased the load on the reworked sediment filling the scour, the resulting re-consolidation of the fill triggered the overlying faulting and collapse above the scour.

Modern iceberg scour morphology seen in the Arctic and Antarctic is consistent with the character of the buried scour in this study. Several studies note the linear nature and plowed berm features on ocean floor surveys that are seen and described here.

-Has the remobilization of the sediments influenced the development of the ring-like structures on the floor of Lake Superior?

The origin of the ring-shaped depressions on the floor of Lake Superior is not well understood. Although similar structures have been described elsewhere in the Great Lakes (Lake Michigan (Berkson et al., 1975; Colman et al., 1994) and Lake Huron (Moore et al., 1994) and Placentia Bay, Newfoundland (Parrott, 1996)), Lake Superior has the most extensive and widely developed system of lake floor rings.

As previously noted, these features were first described by Berkson and Clay in 1972. They proposed that they were the product of syneresis, the spontaneous expulsion of excess pore water by over-saturated fine grained sediment.

In 1984, Flood and Johnson noted the rings while conducting a side-scan sonar survey near the study area. They suggested the possibility that the rings were formed by dewatering of glacial sediments producing a circular vent pattern and conical fractures. The rings were described as a possible stable form of a pockmark produced by escaping fluid.

Cartwright et al. (2004) explored the possibility that the lake floor rings are the result of an immature polygonal fault system (PFS). By making comparisons between the rings and their underlying structures with established

PFS, similarities arose. PFS are features that consist of small-scale, layer bound extensional faulting in a polygonal pattern. Generally occurring in fine-grained sedimentary systems, PFS develop in the early stages of burial without tectonic influence. Concentrations of faulting and rings correlate throughout western Lake Superior, suggesting that one relies upon the other.

Several hypotheses for the origin of PFS have been proposed, these include: density inversion (Henriett et al., 1989); normal gravitational loading of weak sediment (Goulty, 2001, 2002, 2005); and syneresis (Cartwright & Dewhurst, 1998; Dewhurst et al., 1999). Cartwright (2003) in a review of the mechanisms proposed to explain the genesis of PFS argues that origin of these features is open to debate due to the fact that most PFS research has been conducted on deeply buried systems that have little constraint on in situ properties.

Multibeam sonar surveying in the Isle Royale Trough between Thunder Bay and Isle Royale in western Lake Superior by Wattrus et al (2005) identified a well developed field of lake floor rings. The high-resolution data revealed that the rings were actually composed of chains of pockmarks which they refer to as "ring of pearls" structures. These pockmarks mark the location where fluid has escaped the lake-floor. Given the lack of source material for generating gas in the basin this fluid is most likely pore water. The data from this survey also documents a progressive change in the organization of the rings. In the southwest part of the survey, the rings are organized into an irregular network of connected rings. Traced northeast along the axis of the trough, there is a progressive change, first the network of rings is replaced by isolated rings and then finally at the northeastern end of the trough, only isolated pockmarks occur. This change in the character of the lake-floor features suggests that there are changes in the properties of the lake floor sediments that influence the genesis of these features.

The evidence clearly indicates that the rings are produced by fluid escaping from the lake floor. The high-resolution seismic data shows that the rings are underlain by scour complexes consisting of near surface faults developed above paleo-iceberg scour. In order to understand the relationships between these features we need to consider the hydrological history of the basin during the deglaciation of the basin. At this time the percentage of the basin's water supply linked to the LIS would have been significant. This would have been derived from the ice sheet itself and from water diverted along the ice margin. The loss of this water supply would have had a major impact on the hydrologic budget of the basin and it likely led to a significant and probably rapid, drop in lake level. This, coupled with the low permeability of the very fine grained and probably overpresured glaciolacustrine clays, make it likely that the water saturated sediments began to de-water. The scour complexes, in particular the faults created by sediments collapsing into the scours as the result of re-compaction, acted as conduits for the expressed water, carrying it to the surface where it produced the ring-shaped pockmark complexes.

Chapter 6: Conclusions

High resolution seismic data, along with piston cores provide evidence to show that Lake Superior's glaciolacustrine sediments have been subjected to post-depositional deformation. The closely spaced nature of the seismic survey makes it possible to infer a "pseudo-3D" interpretation of the study area.

Within the seismic profile, localized regions of acoustical blanking that are frequently "U" or "V"-shaped are interpreted as iceberg scour. When viewed in plan-view, they exhibit a curvilinear appearance, frequently running for several kilometers across the survey area. Additional support for this iceberg scour interpretation is the presence of localized stratigraphic unconformities and raised lateral berms, caused by iceberg-induced erosion and deformation of the ancient lake floor.

Normal faulting within the sediment overlying the scour is most likely the result of collapse of sediment into scour troughs as the deformed sediment became reconsolidated. When mapping the fault traces across the survey area they exhibit a correlative curvilinear path of underlying iceberg scour. This supports the interpretation of collapse into reconsolidating sediment versus collapse into melting dead ice anchored in the sediment.

There appears to be a connection between the sub-surface post-depositional deformation features and the lake floor ring features. Formation of the ring features has been theorized as stable pockmark features due to conical dewatering and also as an immature polygonal fault system whose genesis is uncertain. A new mechanism linked to dewatering of the sediments triggered by rapid lake level drop associated with the termination of the hydrological linkage between the LIS and the basin is proposed. In this scenario, the scour complexes (the paleo-scour features and the overlying faulted sediments), act as conduits for the expressed pore water that were released as a result of the lake level drop. Recent work within the lake basin has shown that the rings are actually an assemblage of pockmarks referred to as a "ring of pearls." Anomalous amplitude changes in high frequency CHIRP seismic data appear to show fluid evacuation pathways, further supporting the theory that the rings are the result of sediment dewatering. Because of the rings' stratigraphic location, it is likely that the dewatering occurred approximately 9.3 ka, at a time when the LIS lost communication with the basin and lake level dropped.

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