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**ICE-MARGINAL SEDIMENTATION AND PROCESSES OF
DIAMICTON DEPOSITION IN LARGE PROGLACIAL LAKES, LAKE
ERIE, ONTARIO, CANADA**

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Abstract

Detailed studies of coastal cliff exposures through two end moraines form the basis for a model of ice-marginal sedimentation in large ice-contact glacier-fed lakes. The input to the ice marginal environment directly from the glacier included subglacial till and subaquatic flow tills. The subaquatic flow till (thinly-bedded diamicton) was deposited in an apron (up to 1 km wide) along the ice margin. An upward gradient of pore water pressure immediately beyond the ice margin, causing heaving and dilation of the sediments, initiated debris flows of glacially-derived debris (subaquatic flow tills).

Most of the stratified sediments in the ice-marginal zone entered the lake by way of a large proglacial stream. Sedimentation was dominated by quasi- or near-continuous density underflows that resulted in the deposition of a sequence of thick rhythmites.

The glacier in the Lake Erie Basin most likely behaved like an ice stream, with its movement controlled predominantly by a deforming bed of glacial debris, separating the glacier sole from underlying pre-deposited sediments. The deforming bed is preserved as a massive diamicton layer, interpreted here as subglacially-deposited till.

29 **Introduction**

30

31 Studies at the margins of modern glaciers in ice-contact glacier-fed lakes are hampered by the
32 inherent dangers of the environment, while studies on Pleistocene or older deposits in glaciated
33 basins, like the Great Lakes, are far removed in time from the active processes and conditions
34 that prevailed during deposition.

35

36 Results from studies on predominantly Pleistocene mud-rich sediment sequences in the Great
37 Lakes basins, have resulted in numerous hypotheses as to depositional setting and dominant
38 processes of sediment deposition. Hypotheses on the deposition of mud-rich massive
39 diamictos range from subglacial deposition of an advancing or oscillating glacier (e.g. Karrow
40 1967, Barnett 1992) to deposition solely by lake processes (e.g. Eyles and Eyles 1983). Other
41 hypotheses include deposition by surging (Clayton et al. 1985), a deforming bed (Meyer and
42 Eyles 2007, Eyles et al. 2011, Maclachlan and Eyles 2011) and at the grounding line of a
43 floating glacier at the margin of a subglacial lake (Sharpe and Russell 2016).

44

45 The objective of this paper is to describe and interpret mud-rich sedimentary sequences
46 deposited within the glaciated Lake Erie basin of the Great Lakes. The sediment sequences
47 studied provide rare insight as to ice-marginal sediment architecture and depositional process
48 associated with ice-marginal or grounding line deposits formed in large ice-contact glacier-fed
49 lakes. Emphasis is place on the massive diamicton facies, although all facies of the sediment
50 sequences are important in interpreting the environment and processes of deposition.

51

The paper focuses on lake bluff exposures across two key geomorphic features: the Lakeview moraine (Barnett 1985, 1987, 1993) located near McConnell's Nursery, approximately 8 km west of Port Burwell, and the Galt moraine (Taylor 1913) near Sand Hill Park, 13 km east of Port Burwell, Ontario (Figure 1). In both cases lake bluff exposures are nearly continuous, exceeding 1.2 km long and 25 m in height.

Methods

Geological mapping and stratigraphic studies were undertaken along the central shoreline of Lake Erie in response to high lake level and high rates of shoreline erosion. Techniques included the examination of the natural and man-made exposures and the use of soil probes, hand augers and shovels in areas where natural exposures were lacking. Air photos were used to delineate the surface distribution of the geological materials and landforms encountered. Particular attention was paid to the stratigraphy and sedimentology of the sediments exposed by erosion along the Lake Erie shoreline. Exposure was excellent and observations made at over 130 sites along approximately 65 km of active shoreline. Information was collected on sediment description and relationships, structural features and paleocurrent measurements were collected where structures were available (50 measurements per bed). Particle-size analysis was done on multiple samples of the various geological units observed (Barnett 1987, 1993).

Regional Constraints

The Lake Erie basin is the shallowest of the Great Lake basins. It consists of three sub-basins, with an eastern basin, 63 m deep, a central basin about 25 m deep and the western basin, about 10 m deep. The deepest part of the eastern basin is very narrow and is oriented generally east-west. Separating the eastern and central basins is a shallow area less than 15 m deep that consists of two separate ridges referred to as the Clear Creek Ridge and Long Point-Erie Ridge (https://www.ngdc.noaa.gov/mgg/greatlakes/lakeerie_cdrom/html/e_gmorph.htm). The bedrock basin in which Lake Erie is located is much deeper. A long, linear trench about 150 km long, 12 km wide and about 150 m deep (Gao 2011) occurs beneath the central and eastern basins. This trench is filled with glacial and post-glacial sediments (Coakley 1985), most of which were deposited prior to the deposits and landforms discussed below.

The Lake Erie basin played a major role in determining the types and characteristics of the sediments that were deposited within the basin during the Michigan Subepisode (Late Wisconsinan) or Marine Isotope Stage 2 (Karrow et al. 2000). T.C. Chamberlin (1883, 1888) was the first to recognize the lobate nature of the continental ice margin and how landscape topography beneath the glacier affected the shape of the ice margin and direction of ice flow. Initially, the continental glacier entered the northeastern end of the Lake Erie basin and spread southwestward down the axis of the basin (Erie lobe) until it coalesced with another lobe of the continental glacier flowing southward in the Lake Huron basin (Huron lobe). During deglaciation, the reverse is thought to have occurred; the northeastern end of the Lake Erie basin was the last part to become deglaciated. This early concept is supported by the distribution of moraines throughout the southern Ontario landscape (*see* Chapman and Putnam 1951, Barnett 1985, 1992).

97

98 As the continental glacier entered the Erie basin, the existing north-eastward drainage system
99 in the basin became blocked and water levels began to rise in front of and possibly, beneath, the
100 glacier. In general, the further westward the glacier extended into the basin the higher the water
101 level in the basin became as lower lake outlets became covered. However, because several of
102 these high level lakes occupied portions of both the Lake Huron and Lake Erie basins, the water
103 level also depended on the position of the ice margin or grounding line in the Lake Huron basin
104 (Huron Lobe). Several major outlets were used by glacial lakes in the Erie basin including: the
105 Wabash River, Indiana; the Imlay Channel, Michigan; the Uby Channel, Michigan; the Grand
106 River, Michigan; the Indian River Lowlands, Michigan; the Straits of Mackinac, Michigan; the
107 Syracuse Channels, New York; and the Niagara River (Leverett and Taylor 1915; Hough 1958;
108 Prest 1970; Fullerton 1980; Calkin and Feenstra 1985; Eschman and Karrow 1985).

109

110 In the Lake Erie basin in Ontario, the Michigan Subepisode (Figure 2) is represented by a
111 series of events including the Nissouri (Catfish Creek Till), Erie, Port Bruce (Port Stanley Till),
112 Mackinaw (Wentworth Till) and Port Huron (Halton Till) phases (Karrow et al. 2000). The
113 Nissouri, Port Bruce and Port Huron phases are generally considered periods of glacier advance
114 while the Erie and Mackinaw phases of general ice margin recession or standstill.

115

116 The resulting landscape can be described as an elevated, gently-dipping plain. This plain, a
117 former lake bottom, dips gently toward Lake Erie and ends abruptly at the shore cliffs of Lake
118 Erie some 30 m above lake level. The lake plain is interrupted by a series of positive relief
119 landforms, interpreted as morainic ridges, that mark the former locations of the glacier margin

or grounding line as the glacier margin receded. These morainic ridges can rises up to 23 m above the level of the adjacent plain.

Chapman and Putnam (1951, 1984) included the area of this investigation in their Norfolk sand plain physiographic region, which they suggested was part of a gently sloping fan-shaped body of sand, whose apex occurs in the vicinity of Brantford, Ontario. Deposited as deltaic and nearshore sands and silts into large proglacial lakes in the Erie drainage basin, these sediments were subjected to reworking by the waves of subsequent lower lake levels. On close examination, the plain is gently stepped, where the notches or abandoned shore cliffs of these ancestral lakes left their marks.

The sediments examined in this study were deposited during the Port Bruce and Mackinaw phases. The sediment sequences from two areas within the central basin of Lake Erie are discussed. The first is the McConnell’s Nursery area at the westward termination of massive mud-rich diamicton facies within a mud-rich sediment sequence. The lake bluff sections coincide with the mapped extent of the Lakeview moraine. The second area, at Sand Hill Park, near Jacksonburg, Ontario, the western termination of massive, mud-rich diamicton facies that corresponds to the mapped extent of the Galt moraine. The extension of the Galt moraine across Lake Erie is likely the Clear Creek ridge. The once-postulated extension of the Galt moraine Long Point-Erie ridge or Norfolk moraine (Holcombe et al. 2005), appears to be younger than the sediments to be discussed here.

Local Constraints

McConnell's Nursery area (Lakeview Moraine)

The Lakeview Moraine is thought to mark the limit of a minor re-advance of the Erie Lobe and was formed during the latter part of the Port Bruce Phase (Barnett 1993). Its orientation, slightly oblique to the present Lake Erie shoreline, is similar to the orientation of the other moraines formed earlier by the Erie Lobe during the Port Bruce Phase (Mabee, Courtland, Tillsonburg, Norwich, St. Thomas (*see* Figure 1) and Ingersoll Moraines). It is exposed in the Lake Erie shore bluffs near the small community of Lakeview, after which the moraine was named (Barnett, 1985). The Lakeview Moraine can be traced inland to the community of Vienna, 8.5 kms to the northeast. The Lakeview Moraine probably formed at the glacier margin or grounding line within a large ice-contact glacier-fed lake. The Lakeview moraine likely formed during glacial Lake Maumee IV or prior to the establishment of glacial Lake Arkona (Figure 3a).

Based on mapped shoreline features of glacial Lake Maumee IV (Dreimanis and Barnett, 1985) and Glacial Lake Arkona (Barnett 1985, 1993), water depths at the McConnell's Nursery study site were most likely between 75 and 60 m depending on which of these two lakes existed in the Lake Erie basin during the deposition of the Lakeview moraine. If Lake Maumee IV, the main source of sediment input into the lake from terrestrial sources was likely 70 km to the northeast in the vicinity of Paris Ontario (Cowan 1972, 1975). If formed during the subsequent and lower glacial Lake Arkona, the source of sediment input would have been closer, about 40 kms away to the northeast (Figure 3a).

In general, the sediment sequence in the McConnell's Nursery area that is to be described below is related to the Lakeview Moraine, and is similar to, and representative of, inter-bedded mud-rich diamicton and stratified sediment sequences also associated with the Mabee, Courtland, and Tillsonburg Moraines that are well exposed along the Lake Erie bluffs to the west and inland, to the northwest, in various creek bank exposures (Dalrymple, 1971; Barnett, 1982, 1984, 1993; Stewart, 1982; Dreimanis and Barnett, 1984, 1985).

Sand Hill Park area (Galt Moraine)

The Galt Moraine (Fig. 1) can be traced from north of Orangeville southward to the Lake Erie shore near Jacksonburg, Ontario, a distance of some 160 km (Chapman and Putnam 1951, 1984). It is composed of Wentworth Till and associated stratified sediment. The Wentworth Till varies substantially along the length of this moraine. It is a clayey silt to silty clay containing a minor amount of clasts at the Lake Erie shore and is substantially finer than the Wentworth Till described in the Brantford (Cowan 1975) and Simcoe (Barnett 1978) areas to the north. The change in the texture of the Wentworth Till along the Paris Moraine occurs in a series of steps and probably reflects the incorporation of various textured glaciofluvial and glaciolacustrine sediments into the base of the glacier prior to deposition (Barnett 1993).

The sediments associated with the Galt moraine inter-finger with radiocarbon dated organic-bearing sediments of the Jacksonburg Delta (13,360 ± 440 B.P., BGS-929; Warner and Barnett 1986, Barnett 1993) and would place its deposition within the Mackinaw Phase. Elsewhere, within the Norfolk sand plain detrital organic material within sands have been dated between 14.06 and 14.7 ka ¹⁴C BP (Marich 2014), however, stratigraphic relationships have not been

well established. The organic-bearing sediments occur at depths greater than 25 m below the surface and exceed the depths of identified Wentworth Till (Marich 2014). The sites with organic-bearing sediments are located geographically between the Paris and Galt moraines that are associated with Wentworth Till and the palimpsest extensions of the Courtland, Mabee and Lakeview moraines associated with Port Stanley Till. The dates reported by Marich (2014) may relate to the timing of the formation of the Lakeview moraine that is palimpsest in the area studied by Marich (2014).

During the formation of the Galt moraine, water levels in the Erie basin were likely dropping from the glacial Lake Arkona level (Figure 3b, Figure 2) to that defined by the Jacksonburg delta (Barnett 1985, 1993). Water depth at Sand Hill Park during the deposition of the sediment sequence described below would be less than 60 m and could have been as low as 1 or 2 m. The source of terrestrial sediment inflow could have been up to 30 km north of Sand Hill Park if glacial Lake Arkona existed, however, it was most likely much closer, within a few km of the Park (Figure 3b).

Description of Sediments

McConnell's Nursery area (Lakeview Moraine)

The following description of the sediments begins at the top of the lowermost package of diamictos exposed in the lake bluffs at McConnell's Nursery (Figure 4). This diamicton package is part of an earlier sequence of sediments, similar to that described below, that can be

traced westward along the bluff (Figure 4) where it terminates. It is associated with the formation of the Mabee Moraine (Barnett 1993).

Facies A (thick sandy rhythmic couplets)

Resting directly on the lowest diamicton unit is a rhythmically-bedded fine sand and silt unit, designated informally as facies A (M.N. - McConnell's Nursery) (Figure 4). This unit contains couplets composed generally of ripple cross-laminated or climbing ripple drift fine sand and very coarse silt, that grade upward through fine silt into a clay layer at the top. Within the couplets, grain size generally decreases upward (Figure 4). However, in some cases grain size initially coarsens upward then fines within one couplet.

Overall, couplet thickness ranges from about 1 cm to 2.8 m, generally increasing in thickness upwards. The greatest variation in thickness occurs within the coarse-textured portion of the couplet, which ranges from 0.5 cm to 2.6 m. Total number of couplets observed in any one section in the McConnell's Nursery area was 15.

The fine sand-coarse silt component can contain Type A through to Type B climbing ripple drift laminations (Jopling and Walker, 1968). In some cases, sinusoidal (Jopling and Walker, 1968) or draped laminations (Gustavson et al., 1975) are present. Within the coarse component, irregularly-shaped clasts of clay up to 3 cm in diameter occur. Paleocurrent determinations were done on 8 beds of ripple cross-laminated sands with approximately 50 measurements taken per bed (Barnett 1987, 1993). Paleo-current measurements indicate a fairly uniform direction of flow within couplets and between couplets, and generally indicate flow

toward the south-southwest changing slightly to the southwest up section, approximately parallel to the paleo-margin of the glacier in this area.

Samples taken from within the sand/silt portion have a range in Folk's (1968) Graphic Mean between 3.7 and 5.1, and are moderately to poorly sorted (Folk's (1968) Inclusive Graphic Standard Deviation, 0.71 to 1.26). They generally become more poorly sorted upward. The samples are leptokurtic and fine skewed to strongly fine skewed.

The fine portion of the couplets is either silty clay or clay. It contains thin silt or fine sand laminations which may be ripple cross-laminated. There is up to 80 percent clay-size material in some layers. Scour marks on the upper surface of the clay/silt portion have been observed with clay balls present in the overlying sand/silt layer. Thin beds of brecciated clay also may be present.

The clay/silt component of the couplet is dominant in the lower one metre of facies A (M.N.), above which the sand/silt component is dominant. In places, thin beds of interlaminated silt and clay inter-finger with the sediments of facies A (M.N.). Thin discontinuous diamicton lenses and layers may also be present (Figure 4).

At one locality (LP-85-18, Figure 4) a large lens-shaped mound of highly deformed rhythmically-bedded sediments was observed within one couplet. The dimensions of this mound were approximately 5 m high and 13 m across at the base. The layers of clay, silt, and sand within the mound were brecciated and folded; in places bedding was nearly vertical and

dipping steeply to the north. The silty, very fine sands that occur beneath the mound of highly deformed sediments were deformed to a depth of 0.5 m into flame-like structures with heights of about 0.3 m. In an adjacent section (LP-85-17) the sediments were not deformed and a bed of clay balls and small pebbles occurred at about the same stratigraphic position and appear to be part of the same depositional unit.

The overall shape and distribution of facies A (M.N.) is best described as ribbon-shaped. It reaches a thickness of up to 15 m, pinches out laterally both eastward and westward along the Lake Erie shore displaying an apparent width of about 8 km and may extend for 10's of km up paleo-current direction to the northeast, possibly to the sediment input source more than 50 km away.

Facies B (thin silt/clay couplets)

Overlying facies A (M.N.) is a unit of rhythmically bedded silt and clay. This unit, facies B (M.N.), is composed of couplets that range in thickness between 1 and 10 cm. The coarse portion of the couplets consists of planar, ripple cross-laminated or massive silt or very fine sand. The fine portion is silty clay or clay.

The contact between facies A and B is gradational and is commonly irregular and wavy. At a few localities examined, a massive silt bed up to 0.3 m thick occurs at the base of facies B (M.N.) (Figure 4).

Within the couplets, coarse, sand-sized, and small pebble-sized clasts may be present and deform underlying laminations. Occasionally, brecciated silt and clay layers occur. Brittle deformation, including low angle reverse faults, may also be present in **facies B (M.N.)**. Displacements are usually on the order of millimetres to centimetres, however displacements of 1 to 2 metres have been observed. Non-brittle deformation, such as, folds, are also present.

In the upper portion of facies B (M.N.), there are thin discontinuous clayey-silt diamicton layers up to 20 cm thick. Interbedded and gradational contacts with overlying facies C (M.N.) are common.

Facies B (M.N.) varies in thickness from 0 to 10 m, generally thickens westward, and replaces facies A (M.N.) where facies A pinches out laterally about 4 km west of the McConnell's Nursery area (Figure 4).

Facies C (stratified diamicton)

Facies C (M.N.), stratified diamicton, overlies facies B (M.N.) at McConnell's Nursery. This facies is composed of discontinuous layers and lenses of massive clay-silt to silt containing very coarse sand and small pebbles and clasts up to boulder size. These matrix-supported diamicton layers are separated by thin discontinuous fine sand or silt laminations. The layers of massive diamicton range from 20 to 60 cm thick. Their lower contacts can be conformable or erosional as indicated by truncation of the upper part of the fine sand and silt laminations. Brecciation of these intervening silts and sands is common.

Texturally, the massive diamicton layers are similar to subglacially deposited Port Stanley Till in the study area (Barnett 1987, 1993). Cumulative weight percent probability plots of several samples from facies C (M.N.) are presented in Figure 5. The low slope of the cumulative percent probability plots demonstrates the poorly sorted nature of this sediment and is similar to the curve of debris flows presented by Glaister and Nelson (1974). Brittle deformation structures, shear planes, may also be found within facies C (M.N.).

The thickness of facies C (M.N.) ranges between 0 and 8 m. Individual bed thicknesses tend to decrease westward and the unit does not appear to extend much further than 1 km beyond the extent of facies D (M.N.). In places, the diamicton beds form positive-relief mounds about 1 to 2 m high (Figure 6; LP 14 of Figure 4).

Facies D (massive diamicton)

Facies D (M.N.) is texturally similar to facies C (M.N.) but appears structureless or massive. Predominantly a dark greyish brown to greyish brown, clayey-silt to silty clay containing coarse sand grains and a low content of clasts (less than 2 percent), it is a massive matrix-supported diamicton. It usually possesses conchoidal fracture and near vertical joints; sub-horizontal parting and fissility may also be present. Its lower contact can be imperceptibly gradational, sharp, erosional, sheared, or loaded (Figure 7).

This unit ranges in thickness from 0 to 5 m in the McConnell's Nursery area. It appears to thin toward the northwest and eventually becomes inter-fingered with the stratified diamicton of facies C (M.N.) (Figure 4). It can be traced eastward in lake bluff sections for at least 4 km

where it disappears below lake level (*see* Figure 7 in Barnett 1993). The westward termination of facies D (M.N.) roughly corresponds to the position of the distal flank of the Lakeview moraine.

Facies C (M.N.), stratified diamicton, also overlies the massive diamicton layer, facies D (M.N.). Above the massive diamicton, facies C (M.N.) is variable in thickness and can be absent or up to about 3 m thick.

Facies B (M.N.), rhythmically-bedded silt and clay, conformably overlies facies C (M.N.). However, here above the stratified and massive diamicton units (facies C and D) it contains thin discontinuous layers and lenses of massive diamicton and abundant isolated clasts in its lower part that deform underlying sediment. Facies B (M.N.) may also be highly deformed, but involves predominantly non-brittle deformation such as folding.

Sand Hill Park area (Galt Moraine)

The sediment exposed along the Lake Erie bluffs at the Sand Hill Park is summarized graphically in Figure 8. A stratified diamicton layer outcrops occasionally along the base of the bluffs and is related to the formation of the Paris moraine, a moraine formed prior to the Galt moraine that occurs generally west or beyond it, that has been associated with Wentworth till. Figures 9 and 10 display variations in paleoflow direction and particle-size distribution of the sediments exposed along the Lake Erie bluffs between Port Burwell and Jacksonburg, including the area at Sand Hill Park.

Facies A, (thick sandy rhythmic couplets)

Overlying the stratified diamicton of the Paris moraine is a unit of rhythmically-bedded sand, facies A, (S.H.), composed of couplets which generally fine upward from ripple cross-stratified fine sand to silty-clay or clay. Couplets over 10 m thick were observed. The coarse component of the couplets consists predominantly of thick stacked sequences (1 to 2 m) of Type A to Type B climbing ripple drift of fine and very fine sand (Jopling and Walker, 1968). The climbing ripple-drift sequences generally fine upward accompanied by an increase in the angle of climb. Irregularly shaped clay clasts up to 1.5 cm in diameter and detrital organic material including seeds, leaves and twigs, can be present (dated at 13,360+ 440 B.P., BGS-929; Warner and Barnett 1986). The clay component of the couplet is up to 1.5 cm thick and is laterally extensive.

Within the couplets, there are, in some places, several smaller fining upward sequences. Paleo-current measurements indicate strong unidirectional flow within each layer and a gradual shifting upward through the unit from flow toward the west-southwest to flow toward the south-southeast (Figure 9).

Facies A (S.H.) contains lens-shaped inclusions of stratified diamicton, facies B, (S.H.) (Figure 11). The stratified diamictons are composed of individual beds from 5 cm to 1 m thick of massive silty clay containing some very coarse sand, pebbles, cobbles, and occasional boulders. The diamicton beds are usually separated by discontinuous faint laminations of fine sand and/or silt within the lens-shaped inclusions. Several of the massive diamicton layers contain discrete clasts of diamicton, rounded inclusions of rippled sands, and boulders

"floating" or protruding above the diamicton's upper surface. Lower contacts of facies B (S.H.) can be erosional, brecciated, or gradational.

Facies B, (stratified diamicton)

The stratified diamictons, facies B (S.H.), increase in abundance toward the southeast or toward the suggested paleo-ice marginal position of the Galt moraine, where individual massive beds are commonly 0.5 to 1 metre thick and the entire stratified diamicton sequence reaches a thickness of about 15 m (Figure 8, Station L.P. 340). In this area, dish structures occur within the stratified sands and sand dikes penetrate up through some of the diamicton layers.

Facies C, (massive diamicton)

Overlying facies A (S.H.) or facies B (S.H.) at the southeastern end of the Sand Hill Park area is up to 19 m of dark red brown massive silty clay, containing less than 5 percent clasts (facies C (S.H.), Figure 8 and 12). This unit exhibits conchoidal fracture when broken and has a near-vertical set of joints. Conchoidal fracture commonly occurs in massive fine-grained sediments which lack oriented features such as bedding planes or shear planes. Near-vertical jointing is commonly associated with the desiccation of fine grained sediments. The lower contact of facies C (S.H.) with the rhythmically bedded sand unit, facies A, (S.H.) or stratified diamicton unit, facies B, (S.H.) is usually sharp, with or without erosion, incorporation, or deformation of the underlying sediments. Gradational contacts, however, can also be observed at several sites. At one locality, shearing at the lower contact involves the massive diamicton layer itself (Figure 13). The upper contact is gradational with **facies B (S.H.)** where present, and sharp and erosional with facies D (S.H.), often marked by a thin concentration of pebbles.

394

395 ***Facies D, (stratified sand)***

396 Facies D (S.H.) is dominated by flat-bedded or low-angle cross-stratified fine sand. It is well
397 sorted, platykurtic and fine to strongly fine skewed, and the stratification is marked by well-
398 developed heavy mineral concentrations. Occasionally, beds of ripple cross-laminated sand and
399 isolated areas of trough cross-stratified sands occur.

400

401 The trough cross-stratified sand units indicate that flow during formation was generally
402 toward the south. Lenses of stratified diamicton (facies B, S.H.) also occur within facies D
403 (S.H.). Adjacent to the trough cross-stratified sands, truncation of the diamicton beds has
404 occurred and a concentration of pebbles and clasts of diamicton occurs along the bottom of the
405 troughs.

406

407 The upper contact of facies D (S.H.) is marked by an angular unconformity and is overlain by
408 a young paleosol. The paleosol is, in turn, overlain by sands that form the Sand Hill, local cliff
409 top dunes. These sediments are not included or discussed further in this paper.

410

411 **Interpretations**

412

413 **McConnell's Nursery area (Lakeview moraine)**

414 The regional geology and sediment characteristics indicate that deposition of all of the facies
415 exposed at McConnell's Nursery area occurred in an ice-contact glacier-fed lake, either beyond
416 the ice in the lake basin or beneath the glacier. Water depths up against the former ice margin

likely ranged between 60 and 75 m. The former ice margin is marked by the Lakeview moraine (Barnett 1985, 1993).

Facies A (thick sandy rhythmic couplets)

The sedimentary structures and sediment characteristics of facies A (MN) indicate that its overall rhythmicity was probably produced annually (varves). Sedimentation appears to be dominantly from saltation with increasing input from suspension up through the couplet. Deposition of facies A (MN) appears to be dominated by quasi-continuous underflows (Gustavson, 1975; Smith et al., 1982; Smith and Ashley, 1985). The likely large size of the inflowing braided stream network would tend to moderate local variation in flow (direct sedimentation from the glacier) and probably provided a large amount of sediment to the lake.

The high concentration of sediment in the inflowing water ("sediment stratification", Smith and Ashley, 1985) is most likely the main cause responsible for the generation of density underflows. Underflows, controlled by topographic lows within the basin, would tend to flow between previously deposited morainic ridges and/or along the ice margin where, as the ice margin receded into the basin, the deepest water occurred. Paleo-current directions measured were parallel to the paleo-ice margin indicating either very little sediment was being contributed directly from the glacier or the sediment became incorporated into the underflows.

Subsequent underflows (quasi-continuous) may have kept the fine-textured sediments hovering in suspension above the lake bottom (the "nepheloid layer" of Reimer, 1984, in Smith and Ashley, 1985) to be deposited later once underflows ceased.

440

441 Slump-generated surge currents, probably triggered by mass movement further upslope on the
442 delta front, or by mass flows of debris directly from the glacier, may be responsible for the
443 stratification within the fine-grained component of the rhythmite (winter layer) and for the
444 occasional brecciated silt and clay layer. The positive relief mound of deformed rhythmites
445 found within the rhythmite sequence is probably slumped debris and is evidence of mass
446 movement as an active process on the delta slope.

447

448 Erosion on the tops of clay layers and the amount of clay clasts within the summer layers also
449 indicate that underflows were the dominant sedimentary process involved in the deposition of
450 facies A (M.N.).

451

452 Facies A (M.N.) is similar to sediments described by Ashley (1975) in Glacial Lake
453 Hitchcock, which she suggested were deposited in mid-delta foresets (Smith and Ashley, 1985).
454 However, the high amount of clay "balls" and the paleo-geography of the study site suggest that
455 facies A (M.N.) may have been deposited on the lower delta foresets.

456

457 ***Facies B (thin silt/clay couplets)***

458 Facies B (M.N.) inter-fingers with facies A, and is most likely lake bottom sediments (i.e.
459 bottomsets), generally deposited further from the inflowing sediment source. The inter-
460 fingering may be the result of fluctuations in the incoming flow either as lower inflows during
461 several seasons or the periodic abandonment of distributary channels. The presence of ripple
462 cross-laminated fine sand and silt in the summer layers of facies B (M.N.) couplets indicates

that underflows were at least partly involved in the transportation and deposition of the sediments.

Facies C (stratified diamicton)

The interbedded nature of the contact of facies B (M.N.) and facies C (M.N.) and the stratified nature of facies C (M.N.) suggest deposition in water. The textural similarity with facies D (M.N.) and subglacially deposited Port Stanley Till throughout the region (Barnett 1993) indicates a glacial source for facies C (M.N.). Minor erosion along the base of several massive diamicton layers within facies C (M.N.), the silt or fine sand layers between the diamicton layers, the fan-shaped positive relief features, and the interbedded nature with the lacustrine sediments indicate that facies C (M.N.) is composed of subaquatic, glacially derived debris flows or subaquatic flow tills.

The thickness of individual flow tills, between 20 and 60 cm, and their extent, approximately one kilometre beyond the paleo-ice margin defined by the extent of the massive diamicton facies (Facies D, M.N.) in section and the mapped distribution of massive diamicton on surface (Barnett 1993), are comparable to other debris flows or flow tills described elsewhere in the literature. Hester and DuMontelle (1971) discuss a fan-shaped debris flow cone extending approximately three kilometres (2 miles) beyond the Shelbyville Moraine in Illinois. The thickness of till-like material was between 1.2 and 1.3 m. Hartshorn (1958) reported thicknesses of 1.2 to 1.8 m for Pleistocene flow tills and Boulton (1968) reported that flow tills up to 0.75 m thick and extending up to 0.5 km beyond the ice margin occur along the ice margin of Stuebendorfbreen, Vestspitsbergen.

486

487 ***Facies D (massive diamicton)***

488 Facies D (M.N.), the massive diamicton, extends westward in the lake bluff only as far as the
489 Lakeview Moraine, where it becomes inter-fingered with facies C (M.N.). Its lower contact,
490 sometimes erosional and sheared, plus the brittle deformation of the underlying sediments,
491 support the conclusion that facies D (M.N.) is a subglacially-deposited till.

492

493 Facies B (M.N.), above the diamicton units (facies C (M.N.) and D (M.N.)) is a lake-bottom
494 sediment with a decreasing upwards influence of the glacier as indicated by fewer flow tills and
495 less drop clasts (ice rafted debris). Deformation is the result of slumping, buried ice block melt,
496 and/or grounding of ice bergs, or lake ice.

497

498 **Sand Hill Park area (Galt Moraine)**

499 In the Sand Hill Park area, sediments associated with the Galt Moraine are exposed. The
500 sediments associated with the Galt moraine inter-finger with radiocarbon dated organic-bearing
501 sediments of the Jacksonburg delta (13,360+ 440 B.P., BGS-929; Barnett 1993) and would
502 place its deposition within the Mackinaw Phase. Again, the glacier margin supported a large
503 ice-contact glacier-fed lake. Water levels in the Erie basin were likely lower than when the
504 Lakeview moraine formed; less than 60 m and could have been as low as 1 or 2 m. The source
505 of terrestrial sediment inflow could have been up to 30 km north of Sand Hill Park however, it
506 was most likely much closer, within a few km of the Park.

507

508 ***Facies A (thick sandy rhythmic couplets)***

Facies A (S.H.), the rhythmically bedded sand, like facies A (M.N.) at the McConnell's Nursery Site, is probably glaciolacustrine sand, deposited dominantly by quasi-continuous underflows. Facies A (S.H.) however, probably formed closer to the sediment source than facies A (M.N.) at the McConnell's Nursery site, based primarily on grain size and thickness of couplets. The climbing ripple-drift sequences generally fine upward accompanied by an increase in the angle of climb resulting from an increase in the amount of sediment being deposited from suspension relative to the rate of traction sediment transport, indicating waning of the current during deposition (Jopling and Walker, 1968). These sediments are like sediments described by Ashley (Ashley 1975, Gustavson et al., 1975) in Glacial Lake Hitchcock, Massachusetts where sedimentation was suggested to occur on distal prodelta slopes of generally less than 5 degrees. A similar setting is proposed here for the deposition of facies A (S.H.) on the slope of the Jacksonburg delta (Barnett 1985, 1993). The very thick couplets (exceeding 10 m thick, LP 85-5) that occur at Sand Hill Park may indicate that the sedimentation is, in part, related to a subglacial meltwater flood and tunnel channel cutting event that occurred in the area around Lake Simcoe (Barnett 1989, Barnett et al. 1998, Russell et al. 2003). The overflow of meltwater and sediment across the Niagara Escarpment via Guelph area meltwater channels from the tunnel channel cutting event could be recorded in the Lake Erie bluffs.

Facies B (stratified diamicton)

Facies B (S.H.), the stratified diamicton, can be found inter-bedded with facies A (S.H.) and D (S.H.). Its stratified nature, discontinuous lens-shaped distribution with rounded "lobe-like" terminations of layers, the presence of rounded inclusions of stratified sediments, the block inclusions of diamicton and "floating" boulders suggest deposition by debris flow. Dewatering

structures in sand beds near facies B (S.H.) sediments are likely the result of fluidization of the sands during rapid sedimentation and loading by the debris flows from above. The texture of the stratified diamicton is similar to that of the Wentworth Till of the area (Barnett 1993) and facies C (S.H.) and suggests that the debris originated from the glacier or glacially deposited sediments.

Facies C (massive diamicton)

Facies C (S.H.), the massive diamicton, can be traced northwestward only as far as the Galt Moraine where it becomes inter-fingered with facies B (S.H.) sediments. Like facies D (M.N.) at McConnell's Nursery, it is thought that the massive diamicton facies is subglacial till.

Facies D (stratified sand)

Facies D (S.H.), which is really an assemblage of facies, is predominantly the result of deposition in shallow water. The isolated trough cross-beds probably represent the former position of distributary channels on the delta top or front. The heavy mineral concentrations and the symmetrical ripples are evidence of wave activity in this environment (Barnett 1993, Oakes 2002) and may suggest shallowing water depths related to the Jacksonburg delta (Barnett1985, 1993).

Discussion

The sediment sequences described above are summarized in Figure 14. Facies A, B, and C, at McConnell's Nursery and Facies A, B and D at the Sand Hill are interpreted as proglacial lake

deposits. Facies D (M.N.) and facies C (S.H.) are interpreted as subglacial till. Diamicton beds within facies A (M.N.), B (M.N.), C (M.N.), A (S.H.), B (S.H.), and D (S.H.), or the proglacial lake deposits, are interpreted primarily as subaquatic glacially derived debris flows or subaquatic flow tills.

At the McConnell's Nursery area, transitions through facies A, B and C represent the advancing ice margin, facies D (M.N.) overriding of the site by the glacier, and the upper facies C and B, ice margin recession. Similarly, at the Sand Hill, transitions through facies A and B represent an advancing ice margin, facies C overriding of the site by the glacier and facies B and D ice margin recession. Sediment sequences studied and described elsewhere within the glaciated Great Lakes basins exhibit similar facies relationships (Dalrymple, 1971; Barnett, 1982, 1984, 1993; Stewart, 1982; Dreimanis and Barnett, 1984, 1985; Sharpe and Russell 2016).

It should be noted that at this time, water levels were controlled by the uncovering of outlet channels across Michigan and was dependent on the position of the ice margin/grounding line of the glacier lobe in the Lake Huron basin. Marginal fluctuations of the ice lobe in the Lake Erie basin did not affect lake levels. Several factors influenced the type of sediment being deposited within the lake basin, including the amount of sediment entering into the lake from both terrestrial sources and the glacier and the strength and location of density underflow currents.

In general, if ice margin recession was in response to climatic warming or at least greater melting along the ice sheet terminus, then sediment loads entering the lake may have been large and underflows more dynamic; dispersing sediment further out into the lake and into the area of recently uncovered lake bottom along the margin. Possibly during deteriorating conditions, less melting of the glacier would occur, resulting in lower volumes of sediments brought to the lake basin and likely less and weaker underflow activity. The result could produce the fining upward sequence of stratified sediments observed at McConnell's Nursery as the glacier approached. This might be contrary to general thought, but at this site sediment input directly from the glacier appears to have been limited and overwhelmed by sediment entering the lake in underflow currents flowing along the ice margin draining both the glacier margin and adjacent terrestrial surfaces.

Although these two sequences were deposited at different times and in different lakes, they can still be combined in one conceptual depositional model. At the Sand Hill Park area, deposition occurred closer to the inflow source and in shallower water than at McConnell's Nursery (Figure 14). Figure 15 summarizes the possible relationships of these two sites in a three-dimensional block diagram illustrating the resulting sediment sequences, their distribution and the major processes involved.

An excellent review of the processes and sediments of glacier-fed lakes was presented by Smith and Ashley (1985). In this review they suggest that ice-contact glacier-fed lake sediments are characterized by rapid lateral and vertical facies changes (Smith and Ashley, 1985). These changes are the product of the great variability of ice marginal processes. The

presence of "lacustrine fans" (Boothroyd, 1984), or "subaqueous outwash" and their sediments, at the mouth of an ice tunnel introduces coarse-textured water-sorted sediments anywhere along the ice margin. Smith and Ashley (1985, p.203) suggest that this material, containing "syn-depositional primary structures indicating very rapid deposition and post-depositional collapse features produced by melting iceis interbedded with remobilized till (Lawson, 1982; or flowtill, Hartshorn, 1958) and finer grained lake sediments with abundant dropstones derived from floating ice bergs".

During the present study, very little evidence was seen of water-sorted sediment that appeared to be supplied directly by the glacier from ice-tunnels. Alley and others (2003) suggest that where there are subglacial sediments, basal meltwater conduits may be constricted raising basal water pressure, speed and deformation. Glacier input was predominantly by debris flow, producing facies B (S.H.) and facies C (M.N.). Paleoflow measurements in the sands indicated flow generally from the northeast along or parallel to the former ice margin. The very thick couplets (exceeding 10 m thick, LP 85-5) that occur at Sand Hill Park may indicate that the sedimentation is, in part, related to a subglacial meltwater flood and tunnel channel cutting event that occurred in the area around Lake Simcoe (Barnett 1989, Barnett et al. 1998, Russell et al. 2003). However, no evidence was found to support subglacial meltwater flow events down the axis of the Lake Erie basin during the deposition of sediments at the two localities studied contrary to Lewis and Todd (2017).

The large stream which was likely draining exposed land to the north and the glacier margin flowed along the glacier margin and was the primary source of water-sorted sediments. As a

result of this, and the overall size of the lake and the glacier, the high variability in sediments expected in ice-contact glacier-fed lakes was not observed in the sediments of the study area. Locally, however, subaqueous fan sediments may make up a significant proportion of the sediments deposited in the glacial-glaciolacustrine environment (e.g. Boy Scout Gravel Pit, Evenson et al., 1977; or in the Champlain Sea, Rust and Romanelli, 1975).

The lateral transition from massive diamicton into stratified diamicton occurred in both ice marginal settings studied (Figure 4 and Figure 8). In vertical sequence, a corresponding transition from stratified diamicton to massive diamicton back to stratified diamicton occurred (Figure 14). Similar diamicton sequences (Sunnybrook Drift) have been described at the Scarborough Bluffs on Lake Ontario; however, non-glacial processes were suggested for their origin (Eyles and Eyles, 1983). At Scarborough, there is little or no regional control on the geometry of these older sediments or geomorphology and interpretation is based primarily on a two-dimensional view of the sediments, unlike what is available at this study site.

Till Depositional Processes

Subglacial Till

The identification of the massive diamicton layers as subglacial till within glaciated basins is difficult. Characteristics common in terrestrially deposited subglacial tills are often lacking (*see* Boulton, 1976; Dreimanis and Lundquist, 1984; Dreimanis and Schluchter, 1985; Shaw 1985; Dreimanis, 1989 and Evans et al., 2006). It is often the regional distribution of these

layers, their relationships with end moraines and the juxta-position and facies assemblage that make their identification as till possible.

Till deposited sub-glacially in the ice-marginal glacier-fed lake environments studied is usually massive and is often vertically jointed. Low angle shear planes may be visible but are not common. Lower contacts can be sharp, with or without erosion, incorporation, or deformation of the underlying sediments. Sheared lower contacts that involve the overlying massive diamicton layer are rare but provide important clues as to the mechanism of till deposition. Gradational contacts are common with the massive diamicton separated from the underlying sediments by thinly bedded diamicton layers. Load structures can occur along both sharp and gradational contacts. In most cases, the lack of major erosion, incorporation, or deformation is notable and appears common in this ice-marginal glacial-fed lake environment.

Within the study area extensive incorporation of glaciolacustrine sediments occurred prior to till deposition (Barnett 1993). This incorporation probably occurred as a result of freezing onto the base of the glacier in the areas of freezing within a melting zone (Denton and Hughes, 1981) or by extensive deformation of sub-glacial sediments further up-glacial and the formation of a deformable bed. Abundant sediment can be delivered to the glacier by freeze-on as the result of supercooling along an adverse slope or an overdeepening in sediment-floored settings (Alley et al. 2003). Immediately up-ice flow of the study area is the sediment-floored western end of the eastern basin of Lake Erie that is overdeepened (Lewis and Todd, 2017). In addition, “closure of low pressure channels is not uniquely a sign of supercooling, but might arise ... by creep of

soft sediment into channels ...” (Alley et al. 2003, p.139) which may explain the apparent lack of stratified sediment delivered directly to the ice margins studied.

Till matrix properties are generally homogeneous over widespread areas, as has been noted elsewhere in the Great Lakes Region (Karrow, 1976; Kemmis, 1981; Wickham and Johnson, 1981). This homogeneity in matrix properties has been attributed to extensive shearing in debris-rich basal ice prior to deposition, continuous subglacial deformation within the till once deposited and/or by regelation processes, and repeated cycles of erosion, transportation, and deposition of matrix material (Kemmis, 1981). Kemmis (1981) preferred the latter process, repeated regelation, to explain the homogeneity found in Iowa and Illinois tills. Continuous subglacial deformation (deforming bed) was discarded by Kemmis (1981) due to the general lack of evidence of deformation and erosion in sub-till sediments in Iowa and Illinois. Erosion and deformation of sub-till sediments is also not common in the study area, however, subglacial deformation may still have been an important mixing process here.

Extensive shearing in debris-rich basal ice prior to till deposition is not expected to have contributed greatly to mixing of subglacial sediments in the study area. A cold-base marginal zone which enhances shearing within the glacier is not expected to occur when a large body of water is ponded up against the front of the glacier. Warm or wet-based (melted or melting bed) conditions should prevail in this situation. In addition, the glacier would be most likely in a condition of extending flow through the central part of the basin, which would also hinder extensive shearing occurring in the debris-rich basal ice. Extensive freeze-on of subglacial

sediments down flow from an overdeepening (Alley et al. 2003) in the eastern basin of Lake Erie may have supplied abundant sediment to the glacier base.

The ice advance associated with the formation of the Lakeview Moraine extended for about two to three km over a period of about 10 to 15 years. This estimate is based on the distance parallel to ice flow that facies D, (M.N) overlies the pro-glacially-deposited rhythmites of facies A (M.N.) and facies B (M.N.) and the amount of time represented by the rhythmites; assuming the rhythmites are annual. Calculated ice velocities at the glacier sole are therefore about 0.6 m per day. This value is similar to ice velocities obtained by ice streams in Antarctica (Sugden and John, 1976, Bindshadler and Scambos, 1991, Bennett, 2003, Rignot et al., 2011). Ice fluctuations and flow velocities of similar magnitudes are probably reasonable for the Erie Lobe during the formation of several of the other moraines in the study area. The Erie Lobe may be best considered to have behaved like an Antarctic warm-based ice stream. The geometry of and basin depths within the Lake Erie basin (approximately 63 m within the eastern basin) are likely insufficient to float the glacier unless ice thicknesses were very low. However, it may have been possible for a small subglacial lake to have existed in the deepest part of the narrow eastern basin immediately east of the study area.

Surging could have occurred, but the extensive deformation of sediments at the terminus, commonly associated with surged glaciers (Sharp, 1985) was not observed in the study area. Also, surging is often followed by a period of stagnation. Evidence of stagnation (kettle holes, kettle fills), however, is not common in the study area. The Erie Lobe appears to have

remained active and re-advanced several times over relatively short periods of time. The analogy to an Antarctic ice stream seems more appropriate.

Seismic studies and core sampling in Antarctica have revealed the presence of a saturated porous layer beneath an active ice stream (Blankenship et al., 1986, Christoffersen and Tulaczyk, 2003). Subglacial deformation within the saturated porous layer was suggested as the only mechanism for ice stream motion, however near the grounding line "water generated beneath the ice stream may form a film at the ice-till interface, so that both till deformation and water-lubricated sliding may be important" (Alley et al., 1986a, p.58). They suggested that the saturated porous layer beneath the ice stream was composed of deforming till and that deformation caused dilation and porosities of about 40 percent within the deforming sediment. Their analysis suggested that deformation would continue or propagate downward through previously deposited lodged till (or stratified sediments) to the bedrock surface or "until the strain rate became too small to generate stress concentrations large enough to mobilize more till" or stratified sediments (Alley et al., 1986a; Iverson et al. 2007, Iverson 2010). The deformation zone beneath Ice Stream B migrates vertically on a diurnal cycle in response to changing water pressure and supply (Tulaczyk et al 2000a, b).

In the ice marginal glacier-fed lake environment, it is not difficult to imagine the formation of a saturated porous layer of glacial debris beneath the glacier as a result of basal melting. Deformation within this layer could account for the ice flow velocities estimated above, and for keeping the debris dilatant, and for debris mixing, resulting in massive, homogeneous tills.

The models of ice stream-bed motion of Alley et al. (1986a), dependent on till deformation only, or till deformation and water-lubricated sliding, may be applicable in the study area, but with one exception. There is little evidence in the study area for extensive erosion at the base of the till, the possible remnant of the deforming layer. Blankenship et al. (1986) may have erred in relating erosional forms on the bedrock surface to erosion at the base of the 5 m thick deforming layer. The erosional forms on the bedrock surface could have formed prior to the development of the deforming layer, as the glacier ice initially entered the valley. The deforming till layer with its low relief upper surface may simply be filling and smoothing out the irregularities in the bedrock surface and not responsible for the erosion.

Boulton and Jones' (1979) experiment on subglacial till deformation indicated that deformation and movement within the deforming layer decreases downward, almost exponentially. Likewise, the erosion potential of the deforming bed should also decrease downward from the glacier sole-bed interface. If the deforming bed reached a sufficient thickness, erosion of pre-deposited sediments could be minimal or possibly a condition of no erosion at all could exist near the ice margin if the zone of deforming material was within the bed of till (Boulton and Dent, 1974; Boulton et al., 1974).

The abrasion suggested by Boulton and Dent (1974) that can occur within the deforming bed of subglacial sediment in a terrestrial environment may not be significant in the saturated, dilatant, deforming bed environment suggested by Alley et al. (1986a) at the base of Antarctic ice streams or the environment considered here in the ice-contact glacier-fed lake environment.

In addition, pore water pressure build-up within the saturated, dilatant, deforming sediment may prevent extensive abrasion.

Glacially-derived Debris Flows or Flow Till

The identification of the thin beds of diamicton (less than 1.5 m) as glacially derived debris flows or flow tills is often done on the basis of sediment associations. However, several characteristics of glacially-derived debris flows are commonly present, including: conformable lower contacts, or deformation, brecciation and attenuation of the underlying sediments; rounded inclusions, sausage-shaped shears, and hook folds of underlying stratified sediments; fan-shaped accumulation of diamicton layers (flow fans), isoclinal folds of diamicton beds and included stratified sediment layers and "floating" boulders protruding above the upper surface of the debris flow (Dott, 1963, Boulton 1968, Lawson, 1979a, 1981a,b; Hicock et al., 1981; Stewart, 1982, Cohen 1983).

Subaquatic flow tills in the study area are for the most part massive, and are quite similar in texture and composition to the subglacially deposited till (Barnett 1993). Data presented by Cohen (1983) showed a similar relationship of flow tills and subglacial till and this similarity may be the best way of differentiating subaquatic flow tills from other subaquatic mass flow deposits. Their massive nature and the similarity of texture and composition with subglacial till, and hence glacial debris, suggest that the debris flows were highly viscous and comparable to Type 1 sediment flows of Lawson (1979, 1981a, b).

Subaquatic flow tills formed a fairly extensive apron along both paleo-ice margins studied. The interpretation that these subaquatic flow tills originated as debris flows off the front of the ice margin is not totally satisfactory. Accumulating significant amounts of debris on the ice surface prior to flowing is unlikely because ice margins fronting glacial lakes are usually steep with few places where debris can accumulate. As well, up-shearing of debris bands to raise basal debris to the ice surface is probably rare in an ice-contact proglacial lake environment. In addition, the extensional flow suggested above would bring debris toward the glacier sole.

An alternative mechanism for the generation of abundant subaquatic glacially-derived debris flows or flow tills along the ice margin may be the result of a zone where there is an upward gradient of pore water pressure immediately beyond the ice margin (Shoemaker, 1986). Within this zone, soil (sediment) boiling or heaving will occur accompanied by soil (sediment) dilation (Shoemaker, 1986). The potential for the initiation of debris flows in this unstable zone is probably high. Shoemaker (1986) proposed this mechanism to initiate conduits beneath a glacier; however, the fine-grained nature and low shear strength of the subglacial debris in the study areas would probably not allow conduits to be formed or maintained.

The source of material for the debris flows would be subglacially deposited till, or the material from the deforming bed of saturated porous glacial debris beneath the glacier. This would explain the compositional and textural similarities of subaquatic flow tills and subglacial tills in the study area.

This mechanism could also supply the volume of debris flow material needed and explain its distribution, as an apron along the ice margin. Alley et al. (1986b, p.15) have suggested that as a consequence, a deforming till bed "requires deposition of till deltas at the grounding line; available evidence is consistent with the existence of such deltas." The debris flow apron described here (Figure 15) may be a Pleistocene example of the "till delta" predicted by Alley et al (1986b, 1991)

Summary of the Proposed Mechanism of Till Formation

The proposed mechanism for fine-grained till deposition within the study area includes the following steps:

- 1) Extensive incorporation of fine-grained glaciolacustrine sediments by either glaciotectonic deformation and/or by localized freeze-on (Denton and Hughes, 1981, Alley et al. 2003). Freeze-on could possibly be enhanced by either meltwater being forced up-gradient during discharge from the hypothesised sub-glacial lake or supercooling along an adverse slope of an overdeepening (Alley et al. 2003) in the sediment-floored eastern basin of Lake Erie, immediately up-ice flow direction from the areas studied (Lewis and Todd 2017).
- 2) Development of a deforming bed of saturated porous glacial debris by a combination of basal melting, extending flow, overriding of proglacial debris flows, and glaciotectonic deformation of underlying sediments. The first three processes would likely increase in importance closer to the ice margin, while the latter, glaciotectonic deformation, would likely diminish and cease as a contributor of new sediment to the deforming bed. However, this

subglacial deformation would be important in maintaining the deforming bed beneath the glacier sole.

3) Immediately beyond the ice margin, upwelling of porewater due to an upward gradient in porewater pressure forms a zone of instability that can generate debris flows. As the glacier moves forward (advances) these debris flows can become reincorporated into the subglacial deforming bed. However, depending on conditions, not all of these debris flow beds need be destroyed.

4) At the terminus of an ice advance, the debris flow (flow till) apron is preserved and the deforming bed begins to consolidate as subglacial stresses decrease and porewater pressure is allowed to dissipate as the ice stream ablates.

This proposed mechanism of till formation in an ice-contact glacier-fed lake environment would help to explain the till properties, architecture and sediment relationships observed in the study area including:

- 1) The general lack of erosion of pre-existing sediments in the marginal area,
- 2) The massive appearance and vertical jointing within the subglacial till,
- 3) The homogeneity of the subglacial till,
- 4) The sheared lower contacts that involve the till (Figure 11),
- 5) Presence of loaded contacts,
- 6) The common absence of terrestrial subglacial till properties, such as sharp erosional lower contacts, and well-developed shear planes striated clasts, and boulder pavements,
- 7) The common lack of over-consolidation of till deposited in this aqueous environment,
- 8) The large apron of subaquatic flow tills along the ice margin,

846 9) The common stratigraphic relationships of stratified diamicton to massive diamicton back
847 to stratified diamicton often preserved in glaciated basins.

848
849 The thickness of the deforming bed would decrease toward the ice margin because of a
850 reduced ice load; therefore, during a general recession as the glacier thinned, the base of the
851 deforming bed would migrate upwards through the diamicton allowing fabric to be preserved
852 and possibly the development of low angle shear planes within the till.

853
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855
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References

- Alley, R. B. 1991. Deforming-bed origin for southern Laurentide till sheets? *Journal of Glaciology*, **37** (125): 67–76.
- Alley, R.B., Blankenship, D.D., Bentley, C.R. and Rooney, S.T. 1986a. Deformation of till beneath ice stream B, West Antarctica; *Nature*, **322** (3): 57-59.
- Alley, R.B., Blankenship, D.D., Bentley, C.R. and Rooney, S.T. 1986b. Subglacial till and ice stream B, West Antarctica. *In* Abstracts, Chapman Conference on Fast Glacier Flow: Ice Streams, Surging, and Tidewater Glaciers, American Village, British Columbia, p.15.
- Alley, R.B., Lawson, D.E., Evenson, E.B. and Larson, G.L. 2003. Sediment, glaciohydraulic supercooling, and fast glacier flow. *Annals of Glaciology*, **36**: 135-141.
- Ashley, G.M. 1975. Rhythmic sedimentation in glacial Lake Hitchcock, Massachusetts-Connecticut. *In* Glacio-fluvial and Glaciolacustrine Sedimentation, Jopling, A.V. and McDonald, B.C. (eds.), Society of Economic Paleontologists and Mineralogists, Special Publication Number 23, p.304-320.
- Barnett, P.J. 1982. Quaternary geology of the Tillsonburg area, southern Ontario. Ontario Geological Survey, Report 220, 87p.
- Barnett, P.J. 1984. Glacial stratigraphy and sedimentology central north shore area, Lake Erie, Ontario. Geological Association of Canada Annual Meeting, Field Trip Guidebook Number 12, London, Ontario, 42p.

887

888 Barnett, P.J. 1985. Glacial retreat and lake levels, north-central Lake Erie basin, Ontario. *In* Quaternary
889 Evolution of the Great Lakes, *Edited by* Karrow P.F. and Calkin, P.E., Geological Association of
890 Canada Special Paper 30, p.185-194.

891

892 Barnett, P.J. 1987. Quaternary stratigraphy and sedimentology, north-central Lake Erie basin, Ontario.
893 unpublished PhD theses, University of Waterloo, Waterloo, Ontario, 335p.

894

895 Barnett, P.J. 1989. Tunnel valleys in the Georgian Bay area, Ontario; *In* Program with Abstracts,
896 Geological Association of Canada, Montreal, Quebec, p. A89.

897

898 Barnett, P.J. 1992. Quaternary Geology of Ontario. *In* The Geology of Ontario, Ontario Geological
899 Survey, Special Volume 4, pt. 2, Ch. 21, p. 1011-1088.

900

901 Barnett, P.J. 1993. Quaternary geology and stratigraphy of the Long Point-Port Burwell area. Ontario
902 Geological Survey, Open File Report 5873, 212p.

903

904 Barnett, P.J., Sharpe, D.R., Russell, H.A.J., Brennand, T.A., Gorrell, G., Kenny, F.M., and Pugin, A.
905 1998. On the origin of the Oak Ridges Moraine, Canadian Journal of Earth Sciences, v.35, no. 10, p.
906 1152-1167.

907

908 Bennett, M.R., 2003. Ice streams as the arteries of an ice sheet: their mechanics, stability and
909 significance. *Earth-Science Reviews*, **61**(3-4): 309-339.

910

911 Bindshadler, R. A. and Scambos, T.A. 1991. Satellite-Image-Derived Velocity Field of an Antarctic Ice
912 Stream. *Science*, **252**: 242-246.

913

914 Blankenship, D.D., Bentley, C.R., Rooney, S.T., and Alley, R.B. 1986. Seismic measurements reveal a
915 saturated porous layer beneath an active Antarctic ice stream. *Nature*, **322** (3), 54-57.

916

917 Boothroyd, J.C. 1984. Glaciolacustrine and glaciomarine fans: a review. *In* Programs with Abstract,
918 Geological Society of America, Northeastern Section, **16**: p.4.

919

920 Boulton, G.S. 1968. Flowtills and related deposits on some Vestspitsbergen Glaciers. *Journal of*
921 *Glaciology*, **7**: 391-412.

922

923 Boulton, G.S. 1976. A genetic classification of tills and criteria for distinguishing tills of different
924 origins. *In* Till, Its Genesis and Diagenesis, edited by W. Stankowski, Uniwersytet Im. Adama A.
925 Mickiewicza, W. Poznaniu, Seria. Geografia Nr. 12, p.65-80.

926

927 Boulton, G.S., and Dent, D.L. 1974. The nature and rates of post-depositional changes in recently
928 deposited till from south-east Iceland. *Geografiska Annaler*, **56**: 121-134.

929

930 Boulton, G.S., Dent, D.L., and Morris, E.M. 1974. Subglacial shearing and crushing, and the role of
931 water pressures in tills from southeast Iceland. *Geografiska Annaler*, **56A** (3-4): 135-145.

932

933 Boulton, G.S. and Jones, A.S. 1979. Stability of temperate ice caps and ice sheets resting on beds of
934 deformable sediments. *Journal of Glaciology*, **24** (90): 29-43.

935

936 Calkin, P.E. and Feenstra, B.H. 1985. Evolution of the Erie basin Great Lakes. *In* Quaternary Evolution
937 of the Great Lakes, *Edited by* Karrow P.F. and Calkin, P.E., Geological Association of Canada,
938 Special Paper 30, p.149-170.

939

940 Chamberlin, T.C. 1883. Terminal moraine of the second glacial epoch. United States Geological Survey,
941 Third Annual Report 1881-82, p.295-402.

942

943 Chamberlin, T.C. 1888. Rock-scorings of the Great Ice Invasions. United States Geological Survey, 7th
944 Annual Report 1885-86, p.155-248.

945

946 Chapman, L.J. and Putnam, D.F. 1951. The physiography of southern Ontario. University of Toronto
947 Press, Toronto, Ontario, 284p.

948

949 Chapman, L.J. and Putnam, D.F. 1984. The physiography of southern Ontario, 3rd ed.; Ontario
950 Geological Survey, Special Volume 2, 270p.

951

952 Christoffersen, P., and S. Tulaczyk, Response of subglacial sediments to basal freeze-on, 1 Theory and
953 comparison to observations from beneath the West Antarctic Ice Sheet, J. Geophys. Res., 108(B4),
954 2222, doi:10.1029/2002JB001935, 2003.

955

956 Clayton, L., Teller, J. T. and Attig, J. W. 1985. Surging of the southwestern part of the Laurentide Ice
957 Sheet. Boreas, **14**: 235–241. doi: 10.1111.

958

959 Coakley, J.P. 1985. Evolution of Lake Erie based on the postglacial sedimentary record below the Long
960 point, Point Pelee, and Pointe-Aux-Pins forelands. unpublished M.Sc. theses, University of Waterloo,
961 Waterloo, Ontario, 362p.

962

- Cohen, J.M. 1983. Subaquatic mass flows in a high energy ice marginal environment and problems with identification of flow tills. *In* Tills and Related Deposits, *Edited by* Evenson, E.B., Schlacter, C. and Rabassa, J., A.A. Balkema, Rotterdam, Netherlands, p. 255-267.
- Cowan, W.R. 1972. Pleistocene geology of the Brantford area, southern Ontario. Ontario Department of Mines, Industrial Minerals Report 37, 66p.
- Cowan, W.R. 1975. Quaternary geology of the Woodstock area, southern Ontario. Ontario Division of Mines, Geological Report 119, 91p.
- Dalrymple, R.W. 1971. A study of a Late Wisconsin varved sand unit, Port Stanley area, Ontario. B.Sc. Thesis, University of Western Ontario, London, Ontario 140p.
- Denton, G.H. and Hughes, T.J. 1981. The last great ice sheets. John Wiley and Sons, New York, U.S.A., 484p.
- Dott, R.H. Jr. 1963. Dynamics of subaqueous gravity depositional processes. American Association of Petroleum Geologists, Bulletin, **47**(1): 104-128.
- Dreimanis, A. and Barnett, P.J. 1984. Quaternary geology and stratigraphy of the Port Stanley area, Elgin County, Ontario. *In* Summary of Field Work 1984, *Edited by* Wood, J., White, O.L., Barlow, R.B. and Colvine A.C., Ontario Geological Survey, Miscellaneous Paper 119, p.100-101.
- Dreimanis, A. and Barnett, P.J. 1985. Quaternary geology of the Port Stanley area, southern Ontario. Ontario Geological Survey, Preliminary Map P.2827, Geological Series, scale 1:50 000.

- 989 Dreimanis, A. and Lundquist, J. 1984. What should be called till? *Striae*, **20**: 5-10.
990
- 991 Dreimanis, A., and Schluchter, Ch. 1985. Field criteria for the recognition of till or tillite.
992 *Paleogeography, Paleoclimatology, Palaeoecology*, **51**: 7-14.
993
- 994 Dreimanis, A., 1989. Tills, their genetic terminology and classification. *In* Goldthwait, R.P., Matsch,
995 C.L. (Eds.), *Genetic Classification of Glacigenic Deposits*. Balkema, Rotterdam, pp. 17–84.
996
- 997 Eschman, D.F. and Karrow, P.F. 1985. Huron basin glacial lakes: a review. *In* Quaternary Evolution of
998 the Great lakes, *Edited by* Karrow P.F. and Calkin, P.E., Geological Association of Canada Special
999 Paper 30, p.79-93.
1000
- 1001 Evans, D.J.A., Phillips, E.R., Hiemstra, J.F. and Auton, C.A. 2006. Subglacial till: Formation,
1002 sedimentary characteristics and classification. *Earth-Science Reviews*, **78**: 115–176.
1003
- 1004 Evenson, E.B., Dreimanis, A., and Newsome, J.W. 1977. Subaquatic flow tills: a new interpretation for
1005 the genesis of some laminated till deposits. *Boreas*, **6**: 115-133.
1006
- 1007 Eyles, C.H. and Eyles, N. 1983. Sedimentation in a large lake: a reinterpretation of the late Pleistocene
1008 stratigraphy at Scarborough Bluffs, Ontario, Canada. *Geology*, **11**: 146-152.
1009
- 1010 Eyles, N., Eyles, C., Menzies, J., and Boyce, J. 2011. End moraine construction by incremental till
1011 deposition below the Laurentide Ice Sheet: Southern Ontario, Canada. *Boreas*, **40**: 92–104.
1012
- 1013 Folk, R.L. 1968. *Petrology of sedimentary rocks*. Hemphill Publishing Company, Austin, Texas, 170p.
1014

- 1015 Fullerton, D.S. 1980. Preliminary correlation of Post-Erie Interstadial events (16,000-10,000
1016 radiocarbon years before present) central and eastern Great Lakes Region and Hudson, Champlain,
1017 and St. Lawrence Lowlands, United States and Canada. United States Geological Survey Professional
1018 Paper 1089, 52p.
1019
- 1020 Gao, C. 2011. Buried bedrock valleys and glacial and subglacial meltwater erosion in southern Ontario,
1021 Canada; Can. J. Earth Sci. **48**: 801- 818.
1022
- 1023 Glaister, R.P. and Nelson, H.W. 1974. Grain-size distribution an aid in facies identification. Bulletin of
1024 Canadian Petroleum Geology, **22** (3): 203-240.
1025
- 1026 Gustavson, T.C. 1975. Sedimentation and physical limnology in proglacial Malaspina Lake,
1027 southeastern Alaska. *In* Glaciofluvial and Glaciolacustrine Sedimentation, *Edited by* Jopling, A.V.
1028 and McDonald, B.C., Society of Economic Paleontologists and Mineralogists, Special Publication,
1029 Number 23, p.249-263.
1030
- 1031 Gustavson, T.C., Ashley, G.M. and Boothroyd, J.C. 1975. Depositional sequences in glaciolacustrine
1032 deltas. *In* Glaciofluvial and Glaciolacustrine Sedimentation, *Edited by* Jopling, A.V. and McDonald,
1033 B.C., Society of Economic Paleontologists and Mineralogists, Special Publication, Number 23,
1034 p.264-280.
1035
- 1036 Hartshorn, J.H. 1958. Flowtill in southeast Massachusetts. Geological Society of America, Bulletin **69**:
1037 477-482.
1038

- 1039 Hester, H.C., and DuMontelle, P.B. 1971. Pleistocene mudflow along the Shelbyville Moraine front,
1040 Macon County, Illinois. *In* Till/ a Symposium, *Edited by* Goldthwait, R.P., Ohio State University
1041 Press, p.367-382.
- 1042
- 1043 Hicock, S.R., Dreimanis, A., and Broster, B.E. 1981. Submarine flow tills at Victoria, British
1044 Columbia. *Canadian Journal of Earth Sciences*, **18** (1): 71-80.
- 1045
- 1046 Holcombe, T.L., Taylor, L.A., Warren, J.S., Vincent, P.A., Reid, D.F. and Herdendorf, C.E. 2005.
1047 Lake-floor geomorphology of Lake Erie. NOAA, National Geophysical Data Center, World Data
1048 Center A for Marine Geology and Geophysics research publication RP-3, 26p.
- 1049
- 1050 Hough, J.L. 1958. *Geology of the Great Lakes*. University of Illinois Press, Urbana, 313p.
- 1051
- 1052 Iverson, N.R. 2010. Shear resistance and continuity of till at glacier beds: hydrology rules. *Journal of*
1053 *Glaciology*, **56** (200): 1104-1114.
- 1054
- 1055 Iverson, N.R., Hooyer, U.H., Fischer, D., Cohen, P.L., Moore, M., Jackson, G., Kohler, J. 2007. Soft-
1056 bed experiments beneath Engabreen, Norway: regelation infiltration, basal slip and bed deformation.
1057 *Journal of Glaciology*, **53** (182): 323-340.
- 1058
- 1059 Jopling, A.V. and Walker, R.G. 1968. Morphology and origin of ripple-drift cross-lamination, with
1060 examples from the Pleistocene of Massachusetts. *Journal of Sedimentary Petrology*, **38**: 971-984.
- 1061
- 1062 Karrow, P.F. 1967. Pleistocene geology of the Scarborough area. Ontario Department of Mines. Maps
1063 2076, 2077, scale 1:50,000.
- 1064

- 1065 Karrow, P.F. 1976. The texture, mineralogy, and petrography of North American tills. *In* *Glacial Till*,
1066 *Edited by* Legget, R.F., The Royal Society of Canada, Special Publication, Number 12, p.83-98.
1067
- 1068 Karrow, P.F., Dreimanis, A. and Barnett, P.J. 2000. A proposal for revised late Quaternary time
1069 classification in the eastern and northern Great Lakes area. *Quaternary Research*, **54**: 1-12.
1070
- 1071 Kemmis, T.J. 1981. Importance of the regelation process to certain properties of basal tills deposited by
1072 the Laurentian Ice sheet in Iowa and Illinois, U.S.A. *In* *Annals of Glaciology* Volume 2, International
1073 Glaciological Society, Cambridge, England, p.147-152.
1074
- 1075 Lawson, D.E. 1979. Sedimentological analysis of the western terminus region of the Matanuska Glacier,
1076 Alaska. Cold Region Research and Engineering Laboratory (CRREL), Report 79-9, 112p.
1077
- 1078 Lawson, D.E. 1981a. Sedimentological characteristics and classification of deposition processes and
1079 deposits in the glacial environment. Cold Regions Research and Engineering Laboratory (CRREL)
1080 Report 81-27, 16p.
1081
- 1082 Lawson, D.E. 1981b. Distinguishing characteristics of diamictos at the margin of the Matanuska
1083 Glacier, Alaska. *Annals of Glaciology*, Volume 2, International Glaciological Society, Cambridge,
1084 England, p.78-84.
1085
- 1086 Lawson, D.E. 1982. Mobilization, movement and deposition of active subaerial sediment flows,
1087 Matanuska Glacier, Alaska. *Journal of Geology*, **90**: 279-300.
1088
- 1089 Leverett, F., and Taylor, F.B. 1915. The Pleistocene of Indiana and Michigan and the history of the
1090 Great Lakes. United States Geological Survey, Monograph 53, 259p.

1091

1092 Lewis, C F M; Todd, B J. 2017. Linking subglacial meltwater through the eastern Great Lakes, Ontario
1093 and New York (at the time of the Atlantic Heinrich 1 event) with a meltwater flood in the Gulf of
1094 Mexico about 13.5 ka (16.2 cal ka), Geological Survey of Canada, Open File 8181, 1 poster,
1095 doi:10.4095/299649

1096

1097 Maclachlan, J. C., and Eyles, C. H. 2011. Subglacial deforming bed conditions recorded by late
1098 Quaternary sediments exposed in Vineland Quarry, Ontario, Canada. *Sedimentary Geology*, **238**:
1099 277-287.

1100

1101 Marich, A.S. 2014. An assessment of subsurface sediments in the Central Norfolk Sand Plain, Norfolk
1102 and Oxford Counties, Southern Ontario. Ontario Geological Survey, Groundwater Resources Study
1103 14, 132p.

1104

1105 Meyer, P.A. and Eyles, C.H. 2007. Nature and origin of sediments infilling poorly defined bedrock
1106 valleys adjacent to the Niagara Escarpment, southern Ontario, Canada; *Canadian Journal of Earth*
1107 *Sciences*, **44**, 89-105.

1108

1109 Oakes, M.N. 2002. Late Wisconsinian glaciolacustrine sediments deposited in front of a submerged
1110 glacial terminus, Lake Erie Bluffs, Canada. M.Sc. Thesis (Abstract), University of Guelph, Guelph
1111 Ontario, 1p.

1112

1113 Prest, V.K. 1970. Quaternary geology of Canada. *In* *Geology and Economic Minerals of Canada*,
1114 *Economic Geology Report 1*, 5th Ed., p.675-764.

1115

- 1116 Reimer, G.E. 1984. The sedimentology and stratigraphy of the southern basin of glacial Lake Passaic,
1117 New Jersey. M.S. thesis, Rutgers University, New Brunswick, New Jersey.
1118
- 1119 Rignot, E., J. Mouginot, and B. Scheuchl, 2011. Ice Flow of the Antarctic Ice Sheet. *Science*, **333**: 1427-
1120 1430.
1121
- 1122 Russell, H.A., Arnott, R.W.C., Sharpe, D.R., 2003. Evidence for rapid sedimentation in a tunnel
1123 channel, Oak Ridges Moraine, southern Ontario, Canada. *Sediment. Geol.* 160, 33–55.
1124 doi:10.1016/S0037-0738(02)00335-4
1125
- 1126 Rust, B.R. and Romanelli, R. 1975. Late Quaternary subaqueous outwash deposits near Ottawa, Canada.
1127 *In* Glaciofluvial sedimentation, *Edited by* Jopling, A.V. and McDonald, B.C., Society of Economic
1128 Paleontologists and Mineralogists, Special Publication No. 12, p.177-192.
1129
- 1130 Sharp, M. 1985. Sedimentation and stratigraphy at Eyjabakkajökull -An Icelandic surging glacier.
1131 *Quaternary Research*, **24**: 268-284.
1132
- 1133 Sharpe, D.R. and Russell, H.A.J. 2016. A revised depositional setting for Halton sediments in the Oak
1134 Ridges Moraine area, Ontario; *Canadian Journal of Earth Sciences*, **53**: 281-303.
1135
- 1136 Shaw, J. 1985. Subglacial and ice marginal environments. *In* Glacial Sedimentary Environments, *Edited*
1137 *by* Ashley, G.M., Shaw, J. and Smith, N.D., Society of Economic Paleontologists and Mineralogists,
1138 Short Course No.16, p.7-84.
1139
- 1140 Shoemaker, E.M. 1986. Subglacial hydrology for an ice sheet resting on a deformable aquifer. *Journal*
1141 *of Glaciology*, **32** (110): 20-30.

1142

1143 Smith, N.D. and Ashley, G.M. 1985. Proglacial lacustrine environment. *In* Glacial Sedimentary
1144 Environments, *Edited by* Ashley, G.M., Shaw, J. and Smith, N.D., Society of Economic
1145 Paleontologists and Mineralogists, Short Course No. 16, p.135-216.

1146

1147 Smith, N.D., Vendl, M.A. and Kennedy, S.K. 1982. Comparison of sedimentation regimes in four
1148 glacier-fed lakes of western Alberta. *In* Research in Glacial, Glaciofluvial, and Glaciolacustrine
1149 Systems, *Edited by* Davidson-Arnott, R., Nickling, W., and Fahey, B.D., Geo. Books, Norwich,
1150 England, p.203-238.

1151

1152 Stewart, R.A. 1982. Glacial and glaciolacustrine sedimentation in Lake Maumee III near Port Stanley,
1153 southwestern Ontario. Ph.D. thesis, University of Western Ontario, London, Ontario, 348p.

1154

1155 Sugden, D.E. and John, B.S. 1976. Glaciers and landscape: a geomorphological approach. Edward
1156 Arnold (Publishers) Ltd., London, England, 376p.

1157

1158 Taylor, F.B. 1913. The moraine systems of southwestern Ontario. Transactions of the Royal Canadian
1159 Institute, Volume 10, p.57-79.

1160

1161 Tulaczyk, S., Kamb, B., Engelhardt, H.F., 2000a. Basal mechanics of Ice Stream, B. I. Till mechanics.
1162 Journal of Geophysical Research **105**: 463–481.

1163

1164 Tulaczyk, S., Kamb, B., Engelhardt, H.F., 2000b. Basal mechanics of Ice Stream, B: II. Plastic
1165 undrained bed model. Journal of Geophysical Research **105**: 483–494.

1166

- 1167 Warner, B.G. and Barnett, P.J. 1986. Transport, sorting, and reworking of Late Wisconsinan plant
1168 macrofossils from Lake Erie, Canada. *Boreas*, **15**: 323-329.

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FIGURE CAPTIONS

- Figure 1.** Location of end moraines in the study area.
- Figure 2.** Time-space diagram of the Erie Lobe during the Michigan Subepisode (Late Wisconsin).
- Figure 3.** Possible paleogeography constraints to depositional environments: **A)** Lakeview moraine environment of deposition; “A” location of the glacial Lake Maumee IV shoreline and position of main sediment input source. “B” location of glacial Lake Arkona shoreline and position of main sediment input source; **B)** Galt moraine environment of deposition; “A” location of the glacial Lake Arkona shoreline and position of main sediment input source. “B” location of projected Jacksonburg Delta water level shoreline and position of main sediment input source.
- Figure 4.** Sediments exposed along the Lake Erie bluffs near McConnell's Nursery (Lakeview moraine).
- Figure 5.** Cumulative weight percent probability plots of facies C (M.N.) and facies D (M.N.), McConnell's Nursery. **A.** samples of massive diamicton, facies D (M.N.), **B.** samples of stratified diamicton, facies C (M.N.), white area is envelope of massive diamicton samples displayed in A.
- Figure 6.** Stratified diamicton, facies C (M.N.), McConnell's Nursery. A positive relief mound of stratified diamicton (debris flow fan) exposed near LP-14 beyond the extent of facies D (M.N.).
- Figure 7.** Stratified diamicton, facies C (M.N.), McConnell's Nursery overlain by massive diamicton, facies D (M.N.). Note that contact is planar with gentle undulations (white dashed line) and the stratified diamicton beneath is sheared.
- Figure 8.** Sediments exposed along the Lake Erie bluffs near Sand Hill Park (Galt moraine).
- Figure 9.** Paleocurrent measurements along the Lake Erie Bluffs between Port Burwell and Jacksonburg including the Sand Hill Park study area. FA1 – rhythmically-bedded clay with silt, FA2 – rhythmically-bedded silt, FA3 – sand rhythmites, FA4 – low-angle cross-bedded and plane bedded sands and D – diamicton (from Barnett 1987, 1993).
- Figure 10.** Particle-size variation along the Lake Erie Bluffs between Port Burwell and Jacksonburg including the Sand Hill Park study area. FA1 – rhythmically-bedded clay with silt, FA2 – rhythmically-bedded silt, FA3 – sand rhythmites, FA4 – low-angle cross-bedded and plane bedded sands and D – diamicton (from Barnett 1987, 1993).
- Figure 11.** A lens of stratified diamicton, facies B (S.H.) within rhythmically bedded facies A (S.H.) at Sand Hill Park.
- Figure 12.** Massive diamicton facies C (S.H.), Sand Hill Park; as exposed at LP 339, overlying sands and silts of facies A (S.H.). Note low angle shear plane at contact and massive nature and near vertical joints in upper part of photograph.
- Figure 13.** Sheared lower contact (red dashed line) involving the overlying massive diamicton layer indicative of sub-sole deformation.
- Figure 14.** Local sediment sequences for the Lakeview and Galt moraines.

1219 **Figure 15.** Block diagram model illustrating ice-marginal sedimentation in large ice-contact glacier-fed
1220 lake. Processes active include: 1- subglacial melting and formation of the deformable bed of glacial
1221 debris, 2- upward gradient of pore-water inducing heave and dilatancy, initiating debris flows
1222 (subaquatic flow tills), 3- supraglacial flow tills of debris accumulated on the surface of the glacier, 4-
1223 density underflows, and 5- slumping along delta front.

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Figure 5. Cumulative weight percent probability plots of facies C (M.N.) and facies D (M.N.), McConnell's Nursery. A. samples of massive diamicton, facies D (M.N.), B. samples of stratified diamicton, facies C (M.N.), white area is envelope of massive diamicton samples displayed in A.

130x75mm (300 x 300 DPI)

Figure 6. Stratified diamicton, facies C (M.N.), McConnell's Nursery. A positive relief mound of stratified diamicton (debris flow fan) exposed near LP-14 beyond the extent of facies D (M.N.).

87x58mm (300 x 300 DPI)

McCONNELL'S NURSERY SEDIMENT SEQUENCE

SAND HILLS PARK SEDIMENT SEQUENCE

Figure 15. Block diagram model illustrating ice-marginal sedimentation in large ice-contact glacier-fed lake. Processes active include: 1- subglacial melting and formation of the deformable bed of glacial debris, 2- upward gradient of pore-water inducing heave and dilatancy, initiating debris flows (subaquatic flow tills), 3- supraglacial flow tills of debris accumulated on the surface of the glacier, 4- density underflows, and 5- slumping along delta front.

159x140mm (300 x 300 DPI)