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2016

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doi:10.4095/298817

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Recommended citation
Hamblin, A.P., 2016. A Late Ordovician ancient loessite deposit in a glacially-influenced setting?: the Milton member of the Queenston Formation in southern Ontario, and synthesis of the background concepts behind a novel interpretation; Geological Survey of Canada, Open File 7860, 42 p. + figure. doi:10.4095/298817

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ABSTRACT

As part of a larger regional study of Upper Ordovician strata in southern Ontario, 53 outcrops, 10 cores and 120 subsurface geophysical log traces of the Queenston Formation were analyzed. The Queenston Formation, of latest Ordovician age, consists of red non-calcareous siltstone to sandy siltstone with interbedded very fine to fine grained sandstone, and thins over about 250 km from 335 m in the southeast to 22 m in the northwest. The lower Streetsville member comprises a series of stacked thickening- and coarsening-upward shallow marine sequences arranged in an overall coarsening-upward trend, which culminates in the shoreline-related interbedded sandstone and siltstone of the middle Bronte Creek member. These units are overlain by up to 100 m of the upper Milton member, characterized by a thick, fining-upward succession of massive uniform red siltstone which is capped by deep desiccation cracks, a green diagenetic reduction zone and the regionally-extensive subaerial Cherokee Unconformity. In the past, this succession was interpreted as shallow marine to coastal deposits passing upward into supratidal mudflat/sabkha deposits.

In this study, the following facies were identified: 1) red bioturbated mudstone to muddy siltstone (typical of the lower Streetsville member), interpreted as low-energy, shallow marine background deposits, 2) greenish to reddish very fine to fine grained sandstone (interbedded with facies 1 in. coarsening-upward sequences and typical of the lower Streetsville and middle Bronte Creek members), interpreted as higher-energy, nearshore to shoreline traction current deposits, 3) uncommon thin bioclastic calcarenite beds (present in the middle Bronte Creek member), interpreted as higher-energy, nearshore to shoreline traction current deposits, and 4) red, uniform, well sorted, pedogenically-altered siltstone (characteristic of the upper Milton member) and here interpreted as an ancient subaerial loessite deposit in a glacially-influenced setting. New paleocurrent data (175 direct and indirect indicators) suggest a regional shoreline trend of 20°/200°, with a generalized offshore paleoslope direction of 310°.

The Milton member was deposited as an extensive, but rather thin, tabular blanket at ~ 15-20° S paleolatitude near the margin of the Gondwanan continent at the height of a brief, but potent, period of latest Ordovician glaciation. Well sorted, reddened, micaceous silt is the characteristic grain size in thick, massive, unbedded units. These units display uniform blocky/rubbly textures dominated by vertic features, desiccation cracks, fractures, peds and cutans, horizons of caliche nodules and scattered glaebules, gypsum crystals, evaporative crystal molds, weakly-developed calcisols, no fossils or bioturbation, and rare possible rootlets (all interpreted as the result of pedogenic processes). These observations have prompted the proposal of a new sedimentological interpretation for this unit: that of an ancient loessite deposit. If this interpretation is correct, the Milton member of the Queenston Formation represents the first ancient loessite identified in Canada, one of the oldest loessites in the world, and the first anywhere to be associated with the Late Ordovician glacial epoch.

INTRODUCTION

Background

As part of a larger study of Upper Ordovician – Lower Silurian strata of southern Ontario (Hamblin, 1999, 2003, in prep.) a number of outcrops, cores and subsurface geophysical log traces of the Upper Ordovician Queenston Formation were measured and analyzed. Strata of Late Ordovician age are widely distributed in the surface and subsurface of southern Ontario, but are less studied than older and younger units. The sedimentological interpretation of the Queenston Formation has always been enigmatic. This report presents a new interpretation of the sedimentology and depositional setting of part of the Queenston Formation, and a summary of the background literature informing this interpretive concept.

The field area lies between 42°00'N and 45°30'N latitude, and between 79°40'W and 83°00'W longitude; essentially from Mississauga to Windsor in an E-W direction, and from Niagara to Tobermory in a N-S direction. The region is bisected NW-SE by the Niagara Escarpment (providing numerous outcrops) (Figure 1) and there are many oil and gas welbores in the downdip western half of the area (providing subsurface cores and logs) (Figure 2). A total of 53 outcrop measured sections, 10 subsurface cores, and 120 subsurface geophysical log traces, which include the Queenston Formation, were studied (Figures 1 and 2), (see Appendices I and II for location data and Hamblin, 2003 for detailed measured sections). Although many sections only included a few metres of Queenston strata, most have 2-20 m and 8 of the cores have complete sections of the formation, ranging 22-178 m in thickness.

Regional Tectono-Stratigraphic Setting

In southern Ontario, the third-order Upper Ordovician depositional record, Depositional Sequence 3 of Johnson et al. (1992) or Cycle 3 of Sanford (1993) is one of predominantly orogen-derived clastic sedimentation in an overall shallowing-upward trend. The Appalachian margin had become the site of active convergence (primarily through accretion of arc terranes and/or ribbon microcontinents, Van Staal and Hatcher, 2010), transforming the passive margin shelf into an asymmetric subsiding foreland basin (Allegheny Basin) which extended westward as a relatively stable monoclinal ramp onto the craton (Zerrahn, 1978; Sanford, 1993). A major broad inland seaway existed between the Canadian Shield and the active Taconian belt, generally with north-south or northeast-southwest shoreline trends. As the Taconian Orogeny proceeded, the Upper Ordovician clastic wedge prograded outward as the sea withdrew to the west. The entire wedge becomes thinner and younger to the west and northwest.

This sequence comprises a conformable succession of orogen-derived clastics with minor carbonates, deposited in marine to nonmarine environments, with an overall regressive, shallowing-upward trend (Zerrahn, 1978; Lehmann et al., 1994). It sharply overlies the Trenton deepening-upward platform carbonate succession, and is sharply overlain by a regional erosional unconformity at the Ordovician-Silurian boundary (Zerrahn, 1978). The base of the sequence progressively youngs to the northwest, suggesting that tectonic subsidence was the primary control on the stratigraphic succession (Diecchio, 1991; Lehmann et al., 1994). Continued northwestward migration of the Taconian thrust front during this period provided abundant clastic detritus to the deepening/expanding foreland throughout southern Ontario, as far north as Lake Timiskaming (Johnson et al., 1992; Sanford, 1993). Southwestern Ontario occupied the critical paleogeographic transition between the active Allegheny foreland basin and the cratonic platform (Lehmann et al., 1994).

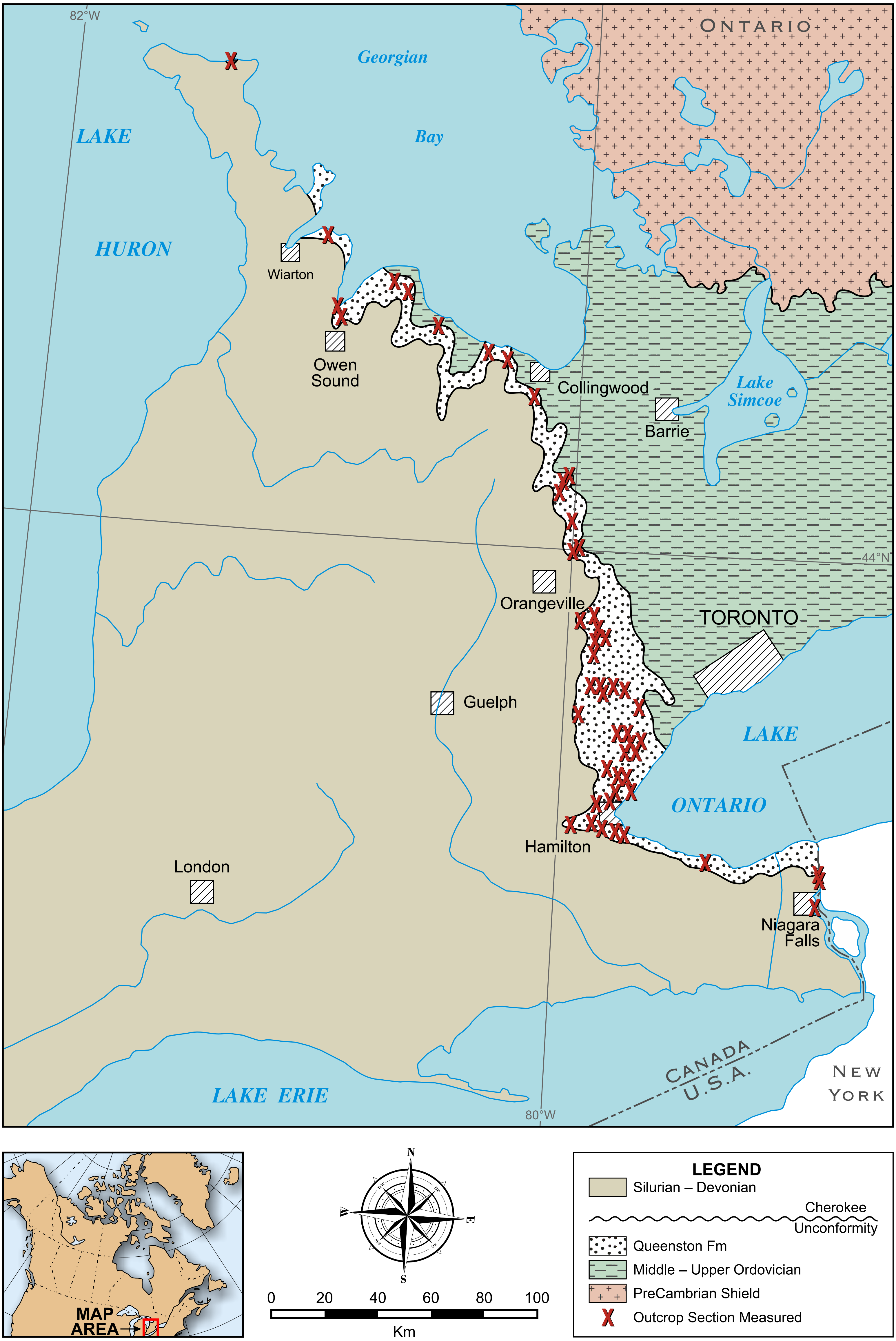

Figure 1. Simplified geological map showing the extent of the Queenston Formation and the outcrop measured section locations used in this study (see Appendix II for location data) (modified from Hamblin, 2003).

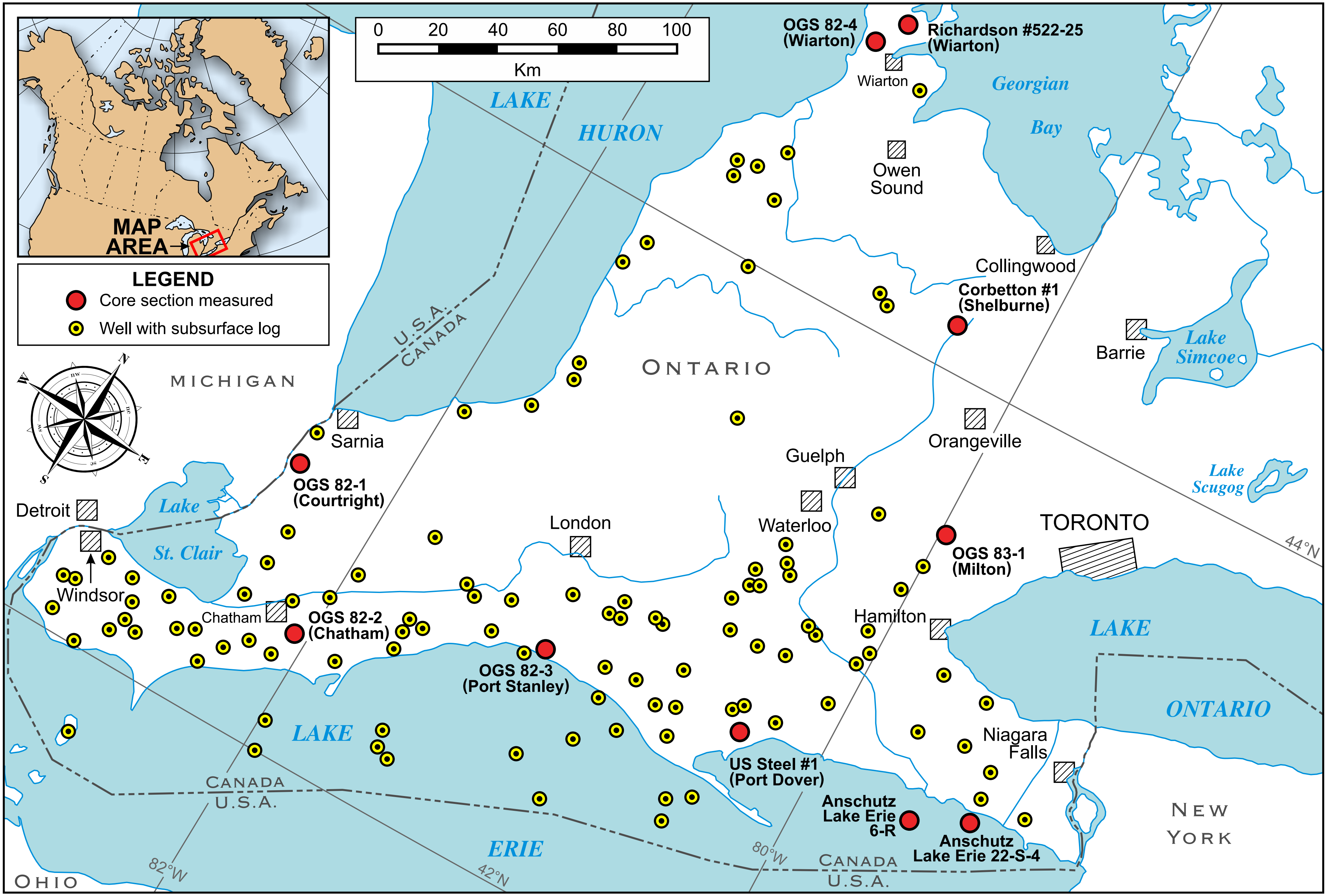

Figure 2. Locations of subsurface well logs utilized and core measured sections (modified from Hamblin, 2003).

In Ontario, the succession (Figure 3) passes from the black, relatively deep water marine shale of the Collingwood Member (of the Lindsay Formation) to the brown and grey shale of the Blue Mountain Formation, to the shallow marine interbedded shale and siltstone of the Georgian Bay Formation, to the marine and nonmarine red siltstone and sandstone of the Queenston Formation (Hutt et al., 1973; Johnson et al., 1992; Sanford, 1993; Armstrong and Carter, 2006; 2010). These units appear to extend to the west and northwest into the area of the flexural peripheral bulge and the intracratonic Michigan Basin (Johnson et al., 1992). Paleocurrent data measured by Zerrahn (1978) in New York and Pennsylvania all indicate flow to the northwest. Castle (2001) detailed four phases of foreland tectonic evolution (with my interpretation of equivalent stratigraphic units in brackets): 1) pre-collisional quiescence, low clastic sediment supply and initial subsidence (Black River-Trenton carbonate platform, 2) orogenic collision and rapid subsidence with sediment starvation (deep-water Martinsburg-Reedsville-Utica shales to the east and Collingwood-Blue Mountain shales to the west in Ontario), 3) late orogenic uplift and abundant clastic sediment supply (Georgian Bay and Queenston progradational wedge in Ontario), and 4) post-collisional waning tectonism (post-Queenston unconformity).

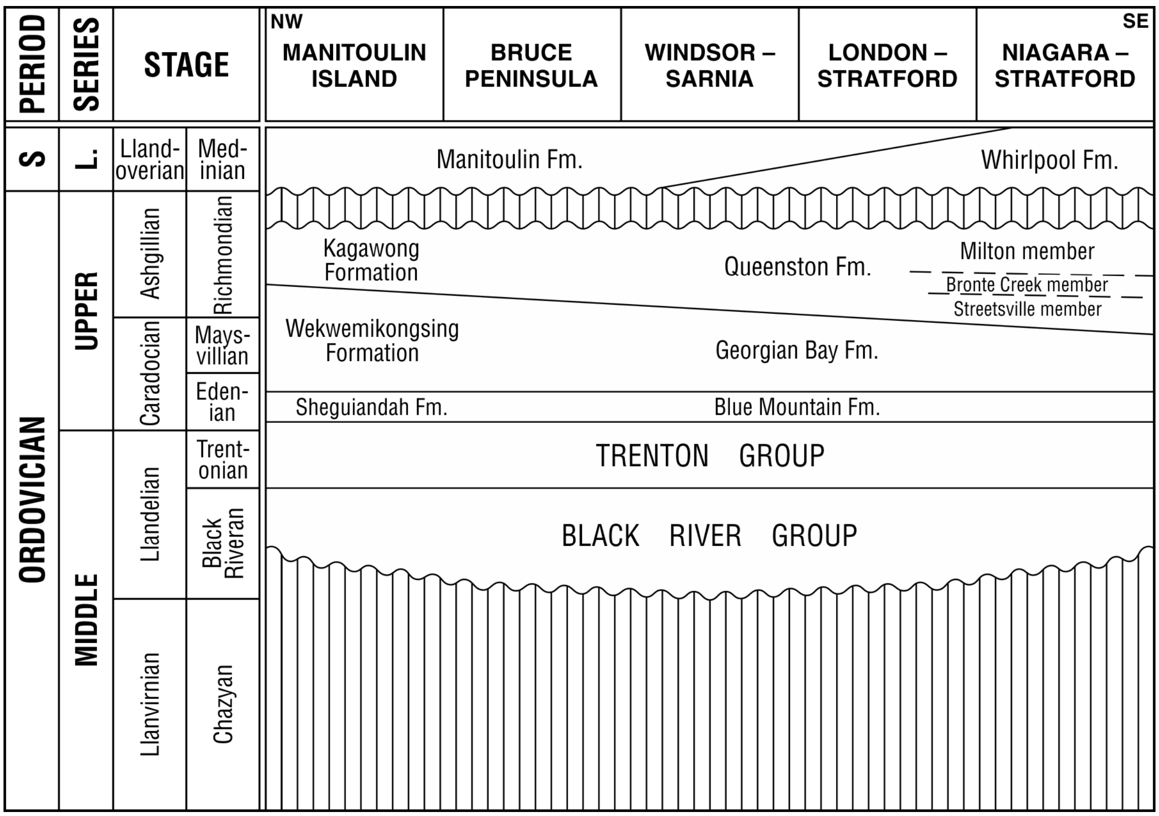

Figure 3. Schematic stratigraphic columns for Middle and Upper Ordovician of southern Ontario (modified from Hamblin, 1999).

The top of the sequence is marked by this subtle disconformity ("Taconic Discontinuity", "Cherokee Unconformity" or "Tuscarora Unconformity"), which coincides with the Ordovician-Silurian boundary (Wheeler, 1963; Sloss, 1988; Dorsch et al., 1994; Armstrong and Carter, 2010). This was interpreted by Lehmann et al. (1994) to represent a combination of effects from glacio-eustatic sea level fall and tectonic flexural tilting to the southeast. The unconformity is unquestionably subaerial and displays regional angularity, suggesting that the erosional hiatus may have been enhanced by tectonic uplift (Lehmann et al., 1994). Even where there is little lithological evidence for significant erosion, there appears to be a faunal hiatus (Middleton, 1987; Rudkin et al., 1998). Middleton (1987) suggested that the latest Ordovician basin filled to just above sea level, after which there was slight erosion of the Queenston Formation and establishment of a northwest-dipping paleoslope by the beginning of the Silurian. According to Johnson et al. (1992) and Rudkin et al. (1998), the unconformity at the top of the Ordovician in southern Ontario (and over much of the mid-continent: Frakes et al., 1993) may relate to a Late Ordovician glacio-eustatic sea level drop. However, Middleton (1987) and Brett et al. (1990) suggested this represents a late Taconian tectonic pulse with an alteration in paleoslope which did not involve any eustatic effects.

THE QUEENSTON FORMATION

Queenston Formation Lithostratigraphy

The name "Queenston Formation" was first proposed by Grabau (1908, 1909), for the strata which occupied the lower 325 m of Hall's original "Medina Sandstone" of New York (Sanford, 1961) with a type locality designated as the Niagara River at Queenston. Sanford (1961) designated a reference well in Crowland Township, Welland County. As a lithostratigraphic unit this distinctive red arenaceous shale is easily recognized throughout southern Ontario, including in a small downthrown fault block in the Ottawa Valley (Wilson, 1964; Johnson et al., 1992). It outcrops all along the base of the Niagara Escarpment from Queenston to Owen Sound and is recognizable in the subsurface to the west (Winder, 1961; Johnson et al., 1992; Armstrong and Carter, 2010). Brogly (1984) suggested the presence of three lithologically-distinct informal members in the Hamilton-Milton area: the lower Streetsville member (60-70 m of interbedded red mudstone and oolitic limestone), the middle Bronte Creek member (30-40 m of red mudstone with abundant siltstone to sandstone beds) and the upper Milton member (50-60 m of dominantly red siltstone to mudstone), the subject of this report.

The Queenston Formation comprises dark maroon to brick red, micaceous and arenaceous, very uniform, thin bedded shale with greenish bands, mottlings and fractures (Liberty, 1955; Caley, 1961; Liberty and Bolton, 1971). The shales are slightly to non-calcareous and very sparsely fossiliferous (Johnson et al., 1992). Thin grey to bluish green, rippled, calcareous siltstones and bioclastic/biostromal limestones represent about 20 % of the rock (Brogly, 1984; Donaldson, 1989) and are especially common near the base and middle, toward the north (Liberty, 1955; Caley, 1961; Liberty and Bolton, 1971; Brogly, 1984; Johnson et al., 1992; Rudkin et al., 1998; Armstrong and Carter, 2010). Donaldson (1989) and Harper (1990) noted a general trend toward upward increase of siltstone/sandstone and calcarenite beds to the middle of the formation, with subsequent decrease upward to the top. To the north, several of these distinctive beds apparently can be traced into the Kagawong Formation of Manitoulin Island (Sanford, 1961; Liberty and Bolton, 1971). These limestone beds contain ooids, brachiopods, ostracods, bryozoans, bivalves and nautiloids (Brogly, 1984; Rudkin et al., 1998; Sharma and Dix, 2004), indicating an Ashgillian (Richmondian) age (Rudkin et al., 1998). The red colour, although very distinctive and the main basis for the original definition of the formation, is likely diagenetic, resulting from oxidation of original grey shale (Liberty, 1955; Liberty and Bolton, 1971). Therefore the definition of the unit as different from the underlying Georgian Bay Formation (based solely on diagenetic colour) is quite arbitrary, accounting for apparent thickness changes over short lateral distances. On the Bruce Peninsula, Armstrong (1988, 1989) found that the red colour is subordinate to grey and green colours, rendering it difficult to differentiate between the Queenston and the Georgian Bay formations. In addition, the greenish bands and fractures result from a second diagenetic bleaching phase of the reddened shale (Caley, 1945). Cross bedding, wave and current ripples, casts of evaporite crystals and desiccation cracks are common (Caley, 1961; Brogly, 1984; Johnson et al., 1992), particularly near the top, where mudcracks filled with the overlying Whirlpool sandstone are present at the upper erosional surface (Caley, 1961). Gypsum, in the form of nodules and thin laminae which may parallel or crosscut bedding are common (Rudkin et al., 1998). The formation thins to the north and west from about 335 m at Lake Erie to about 200 m near Hamilton to 140 m at Halton to 45 m and finally 22 m on Bruce Peninsula (Figure 4) (Liberty, 1955; Sanford, 1961; Brogly, 1984; Johnson et al., 1992).

Armstrong and Carter (2006, 2010) identified the type outcrop locality on the Niagara River at Queenston, with the Highway 403 Roadcut at Ancaster and the Big Bay Roadcut in Bruce Co. as reference outcrops (all included in this study). Excellent cored reference wells are the OGS 82-4 Wiarton core in Bruce County, and the US Steel #1 core in Norfolk County (both included in this study). For geophysical log character, Armstrong and Carter (2006) designated a reference well in Lake Erie, whereas Armstrong and Carter (2010) selected 3 reference wells.

The age of the Queenston has always been difficult to establish due to the paucity of fossils. Foerste (1912, 1916) acknowledged this difficulty, but noted that there were Richmondian (Ashgillian age) fossils in underlying beds and several fossiliferous limestones near the base, similar to those present below, at several locations (confirmed by Caley, 1961). Conodont studies by Tarrant (1977) also suggested an Ashgillian age. From the beginning Grabau (1908, 1909, 1913) was able to physically correlate these strata with the latest Ordovician Juniata Sandstone of New York. Apparently, the youngest Ordovician stage (i.e., Gamachian) is not present, suggesting a significant hiatus, corresponding with the Cherokee Unconformity, before subsequent Lower Silurian deposition (Churcher et al., 1991; Johnson et al., 1992). On Manitoulin Island Foerste (1912) described 25-30 m of grey sparsely-fossiliferous limestones of the Kagawong Formation which conformably overlay the Georgian Bay Formation and are sharply overlain by the Lower Silurian Manitoulin Formation. Foerste (1912) regarded the Kagawong as equivalent to the Queenston of southern Ontario. Copper and Long (1993) pointed out that the uppermost Gamachian stage of the Ordovician is also missing here, and paleokarst features are present, suggesting significant erosion at the top of the Kagawong.

The Queenston Formation is economically important for several reasons. The shales have been extensively and actively utilized for brick and tile clay (Armstrong, 2001). In addition, these strata may have some potential as hydrocarbon reservoirs, and as significant hydrogeological aquifers (Singer et al., 2003). However, the unit is regarded in general as a significant regional aquitard/seal layer (Armstrong and Carter, 2010).

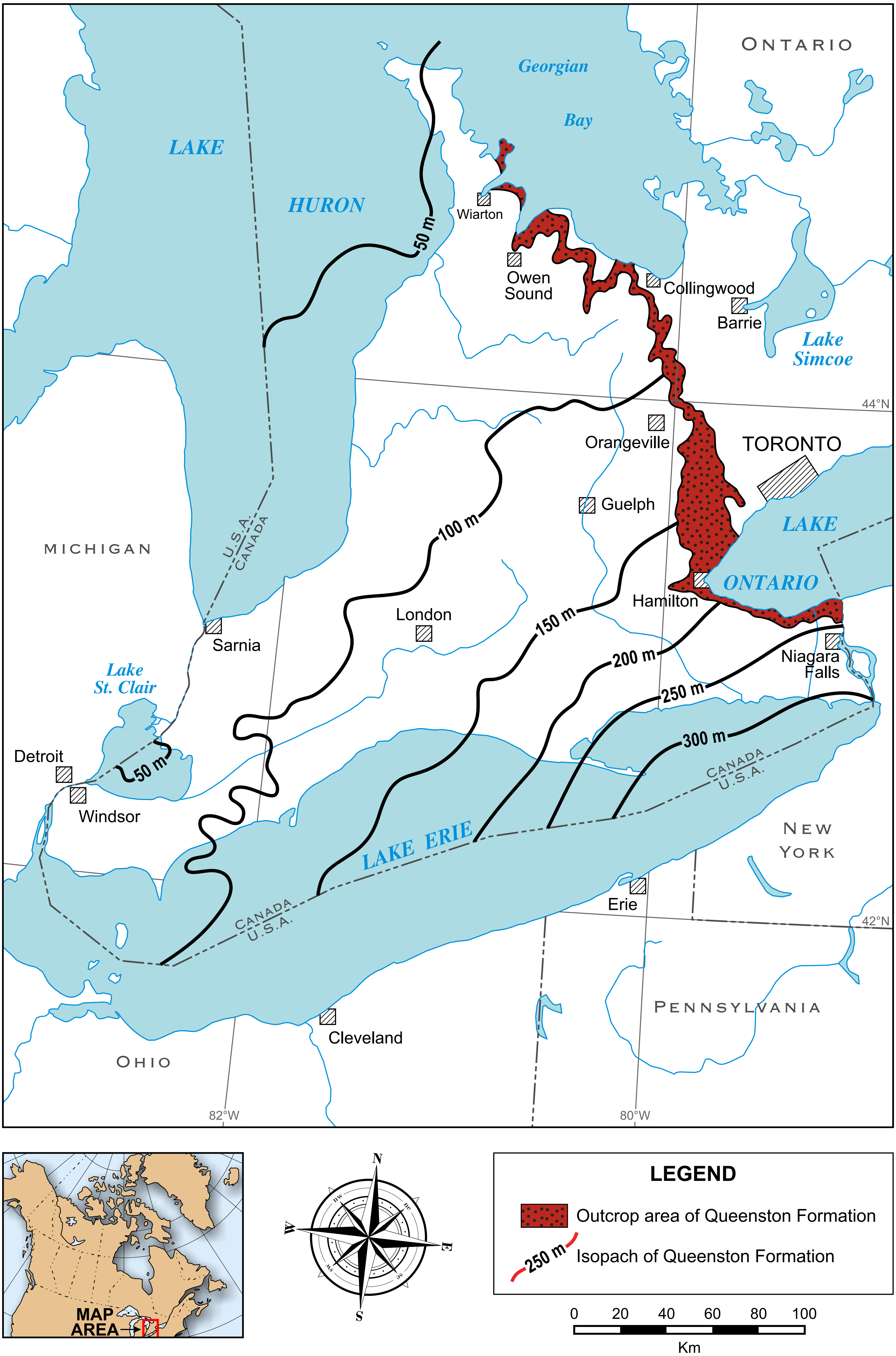

Figure 4. Isopach map of Queenston Formation (modified from Sanford, 1961).

Previous Sedimentological Interpretations and Facies

From the time of its first definition (Grabau, 1909), the Queenston Formation has always been thought of as the distal portions of a large deltaic complex (the "Queenston Delta", lateral equivalent to the Juniata Formation of New York), derived from the Taconian Orogen, but has been the subject of few modern studies. Deposits are considered to be nonmarine to the southeast and marine to the northwest (Johnson et al., 1992). Tarrant (1977) recovered conodonts from a limestone bed in the middle of the formation and pointed out that it intertongues with biostromal reefs to the northwest, suggesting important marine influence from that direction. In southern Ontario, there is no evidence of sandy distributaries, but the rocks do contain graded marine siltstone beds, mudcracks, oolites, evaporites and vertical escape burrows (Brogly, 1984). Brogly (1984) concluded that the Queenston was the result of deposition on a broad supratidal mudflat, the terrigenous equivalent of a coastal sabkha (Middleton, 1987), and that the mud was derived from a river system somewhere to the northeast, now eroded away. Donaldson (1989) described and interpreted several facies associations from one core of the Queenston. These are 1) interlaminated mudstones and bioturbated calcarenites of intertidal mudflat origin, 2) sharp-based graded calcarenites (commonly with lags of intraclasts) with evaporative nodules and cross lamination interbedded with bioturbated mudstone, representing storm beds on a supratidal mudflat, and 3) bioturbated silty mudstone with minor calcarenite deposited in a supratidal setting (Donaldson, 1989). Harper (1990) described four facies, in upward succession as 1) interbedded marine grey sandstone, mudstone and bioclastic limestone, 2) interbedded red bioturbated mudstone and green rippled sandstone with evaporites and mudcracks, 3) interbedded green sandstone and red mudstone with mudcracks, and 4) unbioturbated red mudstone with evaporites (the Milton member of Brogly, 1984, and this report). The last three facies were interpreted to represent deposition on a supratidal mudflat (Harper, 1990).

Brogly *et al.* (1998), working with 7 outcrops and 6 cores in the Hamilton-Brampton area, defined 14 facies in three facies associations, as follows. Continental facies dominate to the southeast and marine facies dominate to the northwest, and they are arranged into two generally regressive successions. 1) Dark grey shale interbedded with thin calcareous siltstone to very fine sandstone and fossiliferous limestone is common in the lower Queenston (and middle part to the north and west), and is very similar to the underlying Georgian Bay Formation. Ripples, flaser bedding, hummocky cross stratification, mudcracks and trace fossils are common. These were interpreted to represent shallow marine deposits with storm-associated graded beds. 2) Red mudcracked shale with interbedded grey shale and thin calcareous siltstone, sandstone and bioclastic limestone typifies the middle portion of the formation. Lags of rip-up conglomerate, gypsum nodules, halite hoppers and trace fossils are common. These sediments were interpreted as coastal deposits affected by storms and fluctuating water levels, and represent a transition between the more marine beds beneath and the more supratidal ones above. 3) Massive red shale with mudcracks, shale troughs, and local lenticular bioclastic siltstone dominates the upper portion of the formation in most outcrops, i.e., the Milton member. These latter strata, the most common in outcrops, were interpreted to represent subaerial muddy coastal plain deposition in a sabkha setting (Gulf of California model) at the continental/nearshore interface (Brogly et al., 1998).

The apparent erosional gap (disconformity, with paleokarst features on Manitoulin Island) at the top of the Queenston is commonly related to the known late Ashgillian glaciation in North Africa and is assumed to represent a major eustatic drawdown (Copper and Long, 1993). This surface is also marked by desiccation cracks filled with overlying sandstone in the Niagara area, and in places, there are rip-ups of Queenston material in the basal part of the overlying Manitoulin Formation (Rudkin et al., 1998), and Whirlpool Formation (Hamblin, 1998).

Previous Paleocurrent Data

Paleocurrent data for the Queenston Formation are extremely sparse, because to date Brogly (1984) and Brogly et al. (1998) are the only modern outcrop-based studies. The former study included 22 measurements of ripple crests from 4 outcrops, which were oriented at an average trend of 119°, and 5 measurements of ripple cross lamination from 2 outcrops, with paleoflow to 222°. This meagre data set implies paleoflow to the southwest from shorelines oriented NNW/SSE (Brogly et al., 1998), but more data are required to make any reasonable interpretations. Zerrahn (1978) reported data (165 measurements) from equivalent strata in western New York, indicating flow of paleochannels to the north or northwest.

QUEENSTON FORMATION FACIES ASSEMBLAGE: OBSERVATIONAL RESULTS

Basic Motif

In the studied sections (see Hamblin, 2003), the Queenston Formation has a minimum thickness of about 22 m (OGS 82-4 Wiarton) and a maximum thickness of 178 m (US Steel DDH #1) (although up to 275 m was reported by Armstrong and Carter, 2010, and up to 335 m by Sharma and Dix, 2004), but averages about 100 m thick, with an obvious thinning trend from southeast to northwest (Figure 4). The Queenston Formation is generally well exposed in the streams west and north of the Toronto region and along the base of the Escarpment in the Niagara region and the Collingwood-Meaford-Owen Sound region. No outcrop exposes the entire thickness, or even more than about 20 metres. Most of the cores used in this study include the entire thickness. It gradationally and conformably overlies the marine grey shales and interbedded thin siltstones/sandstones/limestones of the Georgian Bay Formation, and is sharply and disconformably overlain by the grey medium to coarse grained sandstones of the Lower Silurian Whirlpool Formation in the south, and the greenish mudstones and grey dolostones of the Lower Silurian Manitoulin Formation in the north. In the past, within this coarsening-upward conformable and gradational succession, the lower contact of the Queenston was arbitrarily set where reddish (diagenetic) colours overlying greys of the Georgian Bay began to appear, regardless of lithology, sequencing or depositional style. Strata of the Queenston Formation, studied in 53 outcrops and 10 cores (see measured sections in Hamblin, 2003), consist generally of red, non-calcareous siltstone to sandy siltstone thinly interbedded with very fine to fine grained sandstones. Marine bioturbation and fossils are common at the base, but uncommon to nonexistent through most of the Queenston. The lower portion of the Queenston (Streetsville member) typically displays a series of stacked thickening- and coarsening-upward sequences arranged into an overall coarsening-upward trend (essentially identical to the underlying Georgian Bay Formation, except for the dominant reddish diagenetic colouration) and culminating in the interbedded sandstones and siltstones of the middle Queenston (Bronte Creek member). These are overlain by the upper portion of the Queenston (Milton member) which is characterized by a thick fining-upward succession of massive, uniform red siltstone with very minor thin sandstone interbeds (Figure 5). Interpretation of these latter, rather cryptic, strata has always been difficult, and is the main objective of this report. Sandstone: siltstone ratios range from 1:10 at the base to about 5:1 in the middle, and to 1:20 at the top. The upper contact of the Queenston is very sharp, representing the regional Cherokee Unconformity, and is typically manifest as an irregular surface with complex networks of deep desiccation cracks (Figure 6).

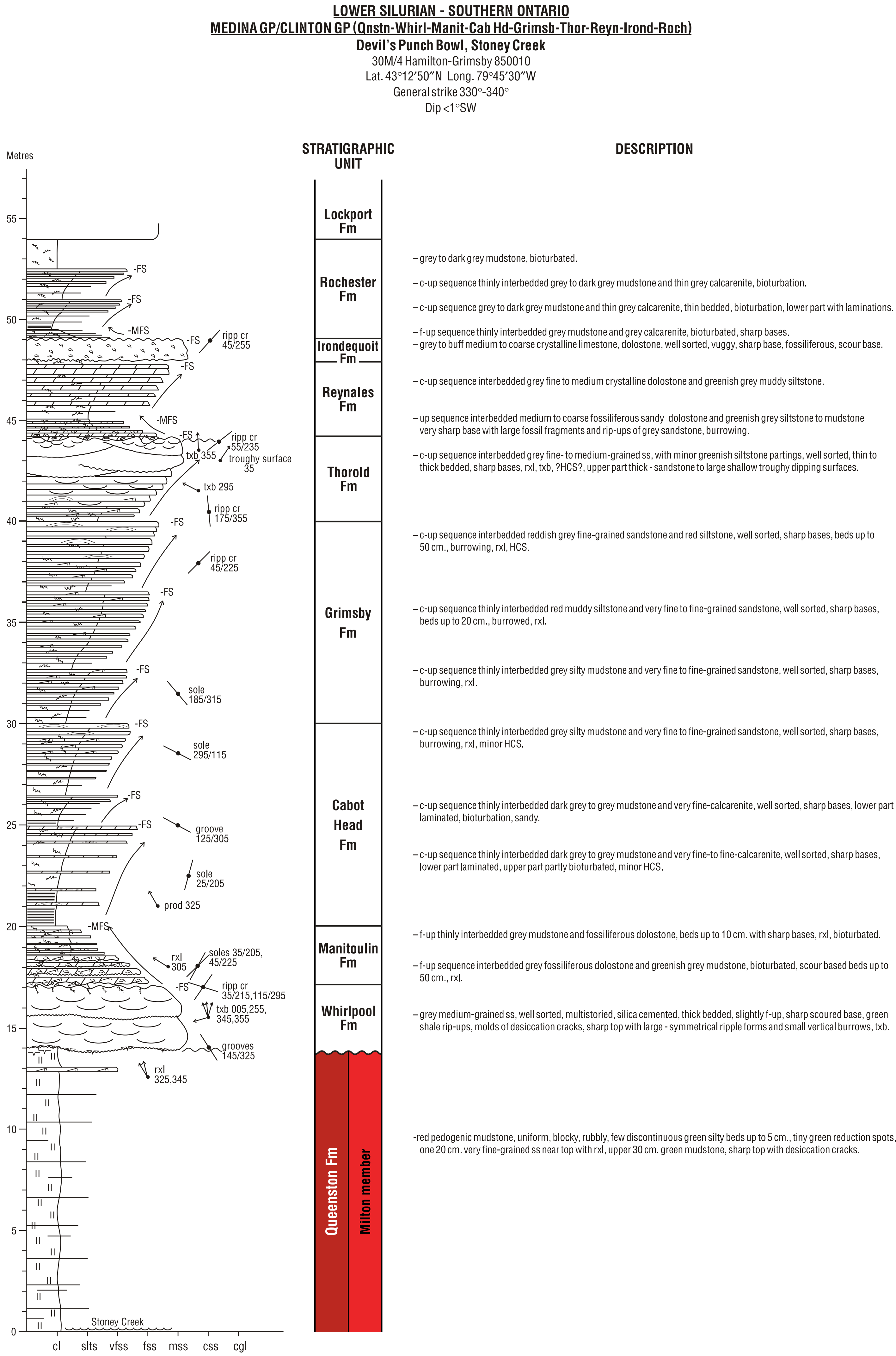

Figure 6. Typical outcrop measured section of Milton member, Devil's Punch Bowl.

Facies 1. Red Bioturbated Mudstone to Muddy Siltstone

Reddish grey to red silty mudstone to muddy siltstone to sandy siltstone is typical of the lower portions of the Queenston Formation (Streetsville member) in many outcrop and core sections. This facies is less volumetrically important to the southeast, but also appears in the middle Queenston in some locations toward the northwest. Poorly to moderately sorted silty mudstone is the most common grain size. Characteristically, it is moderately burrowed to thoroughly bioturbated with numerous, small, vertical and subhorizontal burrows. These deposits are typically massive and uniform, but may display subtle fining- or coarsening-upward, metre-scale trends. This facies rarely occurs in units thicker than a few metres without interbedded "hard bands". In long core sections of the southeast region, bioturbated red mudstone is the common background facies in coarsening-upward sequences typical of the lower 30-40 metres of the Queenston (e.g., Milton, Port Stanley, US Steel #1 cores), and presents similar characteristics to the underlying Georgian Bay Formation mudstones, except for the diagenetic red colouration. Thin discontinuous bioclastic limestone beds and thin to thick greenish to reddish siltstone to fine sandstone beds are commonly interbedded.

These deposits are interpreted to represent low energy background sedimentation in a relatively shallow marine setting.

Facies 2. Greenish to Reddish Grey Very Fine to Fine Grained Sandstone

Reddish to pale greenish to grey "hard bands" are interbedded with the red mudstones and siltstones throughout the Queenston Formation and become increasingly prominent upward toward the middle (Streetsville and Bronte Creek members): they then are less numerous to rare in the upper one-third to one-half of the formation (Milton member). These beds range in thickness from thread-like laminations a few millimetres thick to discontinuous rippled beds 2-20 centimetres thick to significant ledge-like beds up to 60 cm thick (but generally less than 30 cm). In many sections, beds tend to occur in bundles up to several metres in thickness, with individual beds separated by thin red siltstone. Within these bundles, sandstone:siltstone ratios may reach 5:1, whereas the Formation in general has ratios more usually in the 1:10 to 1:1 range. In lithology, these beds include a range of grain sizes from calcareous siltstones to well sorted fine grained, calcareous sandstones to sandy flat-pebble conglomerates. The latter type essentially represents beds which were disrupted/brecciated by severe desiccation crack networks, resulting in the entire thickness occupied by rip-up clasts. The most common grain size in beds of this facies is well-sorted very fine to fine grained sandstone. Individual beds typically have sharp erosive bases and fine upward. Some of the thinner beds, when exposed in larger outcrops, pinch out laterally over tens of metres. The number of siltstone and sandstone beds, thickness and grain size all generally increase upward through the common multi-metre coarsening-upward sequences. Where stratigraphically closely-spaced, beds may be amalgamated into compound beds, separated by wispy threads of dark mudstone. However, individual beds typically have sharp erosive scoured bases with well-preserved sole and tool marks, gutter casts, abundant red or green shale rip-up lags, and burrow casts. Horizontal and low angle lamination, hummocky cross stratification, ripple cross lamination, vertical and horizontal burrows are ubiquitous. A few beds have gypsum nodules and/or molds. Fossil fragments are uncommon, except in thin calcarenite beds which occur in the lower Queenston and in the northwest region, and which are more similar to those of the underlying Georgian Bay Formation. Thicker beds, especially at the tops of multi-metre coarsening-upward sequences, commonly include hummocky cross stratification (e.g., Port Stanley core) and minor trough cross bedding. Common horizontal burrows include *Planolites* and *Chondrites*, whereas vertical burrows include *Skolithos* and ?*Scoyenia*. Bed tops may be bioturbated and gradational, or commonly sharp and rippled with preserved symmetrical and 3-D interference ripples and complex networks of desiccation cracks. A few beds, generally at the tops of coarsening-upward sequences, are characterized by distinct soft sediment deformation, contorted lamination and ball-and-pillow structures.

As discussed in a later section, numerous paleocurrent indicators suggest a regional paleoshoreline trend oriented NNE-SSW, with an offshore direction to the WNW.

This facies is interpreted to represent higher-energy traction current deposition in shallow marine nearshore to shoreline settings. Thick developments of this facies in the middle Bronte Creek member in the southeast are interpreted as a previously-unrecognized NNE/SSW-trending shoreline trend which may have significant hydrocarbon and groundwater potential.

Facies 3. Thin Bioclastic Calcarenite/Limestone to Calcirudite

Greenish grey to reddish grey, argillaceous to sandy, bioclastic limestone beds are infrequently interbedded with the red mudstone and sandstone facies, especially toward the north. A particularly thick development of these beds occurs in the Owen Sound/Bruce Peninsula region (e.g., Sutton Point, Big Bay), where 1-2 m bundles of bioclastic limestone beds are concentrated at the tops of coarsening-upward sequences in the middle Queenston Formation (Bronte Creek member). Here, this facies may be related to the coeval Kagawong Formation of Manitoulin Island, but elsewhere, the facies is uncommon. Individual beds range in thickness from a few centimetres thick to significant ledge-like beds up to 70 cm thick (but generally less than 20 cm). In lithology, these beds include a range of grain sizes and compositions from fine crystalline, to poorly sorted or well sorted medium to coarse crystalline calcarenites, to calcirudites. These beds are commonly composed of sand- to pebble-sized fossil fragments, particularly of brachiopods, crinoids, bryozoans, and pelecypods, set in a calcareous silty or sandy matrix. Oolites are also a common component and may dominate in some beds. In the Richardson #522 core, one 20 cm algal-laminated stromatolitic bed is present near the top of the Queenston. Individual beds typically have sharp erosive scoured bases with mudstone rip-ups and sole marks, although several display the casts of desiccation cracks from the underlying red mudstone. Many beds appear to be massive, but horizontal and low angle lamination, ripple cross lamination, wisps of green mud and vertical and horizontal burrows are present. Bed tops are typically sharp and flat with desiccation cracks, or sharp and rippled with preserved symmetrical and 3-D interference ripples and small burrows. Many beds in certain localities have such a concentration of networks of desiccation cracks that the bed is essentially composed of a flat-pebble conglomerate of disrupted micritic clasts. Many of these beds, when exposed in larger outcrops, pinch out laterally over tens of metres. As discussed in a later section, the few paleocurrent indicators measurable suggest a regional paleoshoreline trend oriented NNE-SSW, with an offshore direction to the WNW.

This accessory facies is interpreted to represent higher-energy traction current deposition in a shallow marine nearshore and/or tidal flat coastal setting associated with shorelines.

Facies 4. Red Uniform Well Sorted Pedogenic Siltstone (the Milton member)

Brick-red to maroon, massive, thick-bedded, uniform, mudstone to silty mudstone to siltstone to sandy siltstone is typical of the upper Queenston (Milton member) and, indeed, is the classic "red shale" facies most researchers associate with the formation (Figure 7). These strata represent the bulk of the stratigraphic unit referred to as the "Milton Member" by Brogly (1984), and may occupy as much as the entire upper two-thirds of the total Queenston thickness, especially to the southeast (see core and outcrop measured sections). Significant units of this facies also occur in the middle part of the Formation, interbedded with thin sandstone facies, but are much less common and thinner in the lower part. Moderately- to well-sorted clay to medium silt is the characteristic grain size and may appear in monotonous units up to tens of metres thick. The deposits are commonly micaceous. In thin section, grain size ranges 10-100 μ, generally 20-60 μ (i.e., medium to coarse silt) of angular to subangular particles, set in a groundmass of red calcareous clay. As a soft, recessive lithology, this facies most commonly occurs in outcrop preserved beneath the resistant beds of the overlying Whirlpool (sandstone) or Manitoulin (limestone) formations, but thick exposures are less common. The cores provide long continuous sections of this facies, and demonstrate that the Milton member generally displays an overall fining-upward trend. Also characteristic is a blocky, rubbly to massive, commonly vertically-fractured macro-texture and a near-complete lack of bedding or lamination. Rather than an overall aspect of horizontal lamination common in most fine grained facies, here the overall sense is of post-depositional vertic structure/texture. Desiccation cracks are ubiquitous in most sections and very common in some. Slickensides and green staining are common on the surfaces of the various scales of primarily vertical fractures.

Figure 7. Milton member uniform red siltstone exposed at "Queenston Badlands", near Inglewood.

In the Port Stanley core and several outcrops, interpreted peds and cutans, which are hallmarks of the soil-forming processes, are well-developed (Figure 8). Small, rounded, greenish-white, calcareous caliche nodules and glaebules, up to 2 cm in diameter, are commonly scattered throughout units of red mudstone, or may appear as discrete calcrete horizons. Tiny, irregular, green reduction spots are ubiquitous throughout the Queenston shales (Figure 9). At the Big Bay section, and in several cores, small to large nodular masses of orange to white gypsum crystals occur on multiple bedding planes within red silty mudstones, and at another section (Red Hill Creek), vuggy molds of evaporitic minerals occur in red mudstone. The Courtright core also displays large pink gypsum nodules in the Milton member. There is no evidence of extensive bioturbation and, to my knowledge, no fossils have been recovered from these fine-grained deposits. However, in one core (US Steel #1) vague, delicate, carbonized vertical root-like structures are present in thick uniform red silty mudstone near the top of the member.

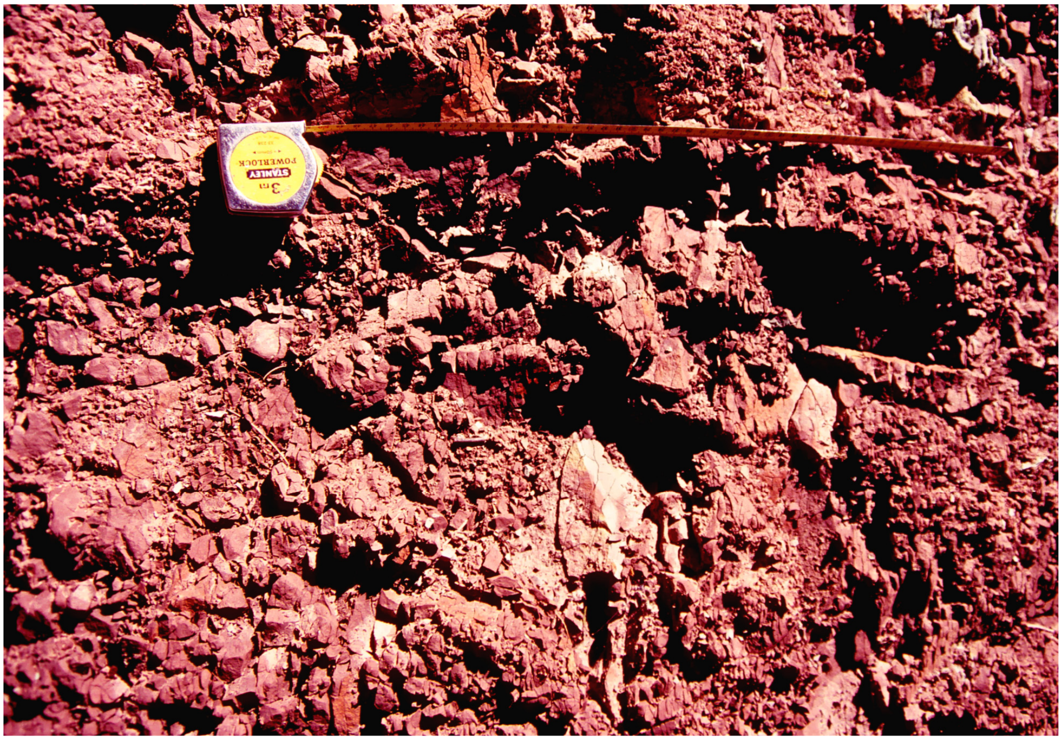

Figure 8. Brick-red pedogenic siltstone with vertical ped structures and slickensided cutan surfaces, Indian Falls, near Owen Sound.

Figure 9. Thin, greenish sandy siltstone bed with sharp base, ripple cross lamination, and desiccation cracks, Sixteen Mile Creek, Lion's Valley Park.

Units of red siltstone are commonly separated by bundles, up to 1 m thick, of thin discontinuous beds of greenish, calcareous, coarser siltstone to very fine sandstone. Individual beds of this lithology are generally sharp-based with mudstone rip-ups, up to 20 cm thick (typically 2-8 cm) and commonly have gradational tops (Figure 9). Small fossil fragments are rarely present in these beds. They may have horizontal lamination, ripple cross lamination and desiccation cracks, but thin beds typically have been altered to a massive state with gradational boundaries. In thick sections of the Milton member, especially toward the south, these siltstone hard bands are less common, but depict an overall fining-upward trend. At the top of one outcrop section (Grindstone Falls) several very thin, white bentonite beds are preserved beneath the Cherokee Unconformity. The upper surface of the Queenston Formation, preserved beneath the unconformable erosional base of the overlying Lower Silurian Whirlpool Formation, commonly preserves a network of deep desiccation cracks filled with sand grains from the Whirlpool and extensive green diagenetic reduction staining in the upper few metres of strata (as at Anschutz Lake Erie 22-S-4).

Although sedimentological interpretation of this cryptic facies has been varied to date (supratidal mudflat according to Donaldson (1989) and Harper (1990), and muddy sabkha according to Brogly et al. (1998)), the rest of this report details a novel, and completely different, alternative interpretation for Facies 4 which dominates the Milton member.

Paleocurrents

Abundant paleocurrent data were obtained from the Queenston (Figure 10). A total of 105 direct paleoflow indicators (104 rxl, 1 txb) from 28 outcrops, presents a generalized depositional paleoslope direction with a mean vector of approximately 310°. Additionally, 70 paleoflow trend indicators (45 ripple crests, 17 sole marks, 7 scours, 1 gutter), with significant scatter, from 12 outcrops, suggest a possible regional shoreline trend of about 20°/200°, and a depositional slope trend of 110°/290°, with the offshore direction being to the WNW. My interpretation of these data is that for the Upper Ordovician, Queenston shoreline-nonmarine depositional system in southwestern Ontario, the shoreline trend of orogen-derived deposits was about NNE/SSW and the offshore dispersal direction was to the WNW. This direction is approximately perpendicular to the Taconian Orogen and toward the craton. These interpretations are similar to conclusions reached by other researchers using similar data sets and/or regional relations (Zerrahn, 1978; Brogly, 1984).

Interpretation of the enigmatic Milton member

Primarily due to its ubiquitous reddish colour, the Queenston Formation has typically been interpreted as dominated by continental deposits through the middle and upper portions (Grabau, 1909; Brogly, 1984). In this study, I largely concur with this generalization, especially for the upper Queenston Milton member above the nearshore-shoreline-related Bronte Creek member. However, the vexing question of what *kind* of continental deposits are represented by the thick red siltstone-dominated strata (Facies 4) of the Milton member has not yet been satisfactorily addressed. Brogly (1984) and Brogly et al. (1998) have provided the most complete and plausible interpretation to date: that of a muddy subaerial coastal plain in a sabkha setting. However, this interpretation might imply that some bioturbation, mudstone beds, marine fossils and channelized sandy units could be expected. None of these are known from this part of the Queenston. Conversely, detailed observations made in this study, combined with a more comprehensive view of the global and regional background controls, suggested to me another more novel, perhaps speculative, interpretation: the Milton member as an ancient loessite deposit in a Late Ordovician glacially-influenced setting. Following is a brief summary of these important background concepts and observations, and the application of these to the Milton member strata.

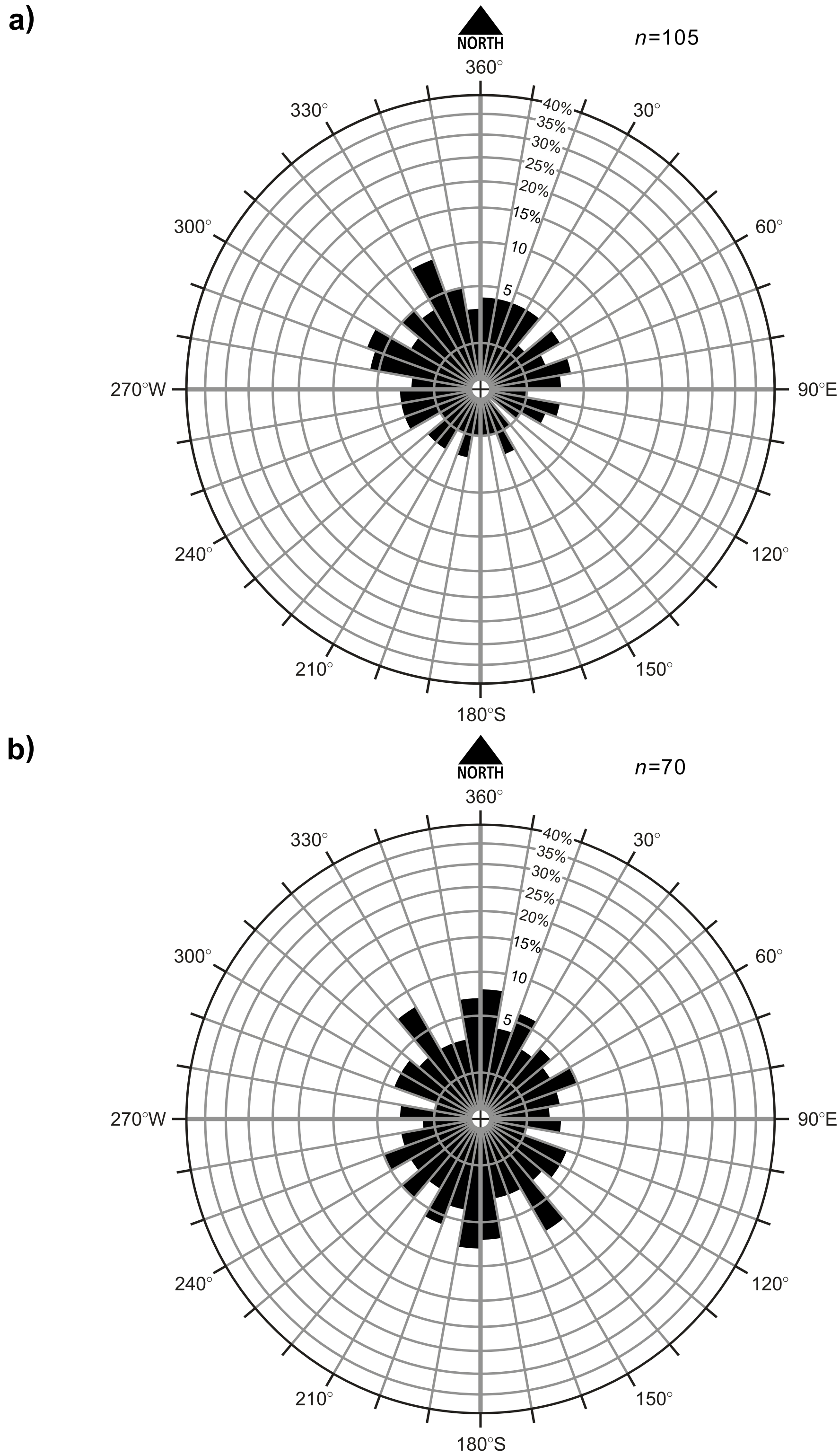

Figure 10. a) Queenston Formation direct paleoflow indicators measured in this study, suggesting depositional paleoslope to 310° (NW), b) Queenston Formation indirect paleoflow indicators measured in this study, suggesting paleoshoreline trend of 20°/200° with offshore direction to the WNW.

THE LATE ORDOVICIAN LANDSCAPE REVOLUTION: BACKGROUND TO NEW INTERPRETIVE CONCEPTS FOR THE MILTON MEMBER

Late Ordovician Glaciation

Climatic Setting

Southwestern Ontario lay at about 15-20° S paleolatitude in Middle Ordovician to Middle Silurian time, on the distal, northwest margin of the Appalachian Foreland Basin, during a phase when the Gondwanan continent was moving over the South Pole (Ziegler et al., 1979; Ziegler, 1981; Scotese and McKerrow, 1991). Laurentia was rotated about 45° clockwise from its present position (Ziegler, 1981; Middleton, 1987; Scotese and McKerrow, 1991), leaving this area in the southern tropical Trade Winds belt, where Late Ordovician winds would blow from east to west across the paleoshorelines (Brogly, 1984). However, the Late Ordovician-Early Silurian Cool Mode (Frakes et al., 1993) was beginning and the Late Ordovician glacial phase, predominantly in the southern hemisphere Gondwanan continental fragments of Africa and South America (Crowell, 1981; Sheehan et al., 1997), began to dominate world climates in the Caradocian (Middleton, 1987; Frakes et al., 1993; Lavoie and Asselin, 1998; Saltzman and Young, 2005). Although glaciation and mass extinction reached their maximum in the final 0.5 to 1.0 M.Y. of the Ordovician (Sheehan et al., 1997), there was a major decrease in faunal diversity in mid Caradocian time which may represent the initial response to the climatic shift from "greenhouse" to "icehouse" conditions (Brookfield and Brett, 1998; Patzkowsky et al., 1997; Page et al., 2007). Concomitant with this climatic shift, a significant global eustatic sea level fall may have occurred, and affected deposition in southern Ontario (Copper and Long, 1993; Rudkin et al., 1998).

There is abundant evidence of Late Ordovician glaciation throughout the Gondwanan landmasses (Berry and Boucot, 1973; Lenz, 1976; Dennison, 1976; Hambrey, 1985; Brenchley, 1988; Brenchley et al, 1991). The major ice centres were sited on the North African craton (Crowell, 1999) but at its height, continental glaciers appear to also have covered parts of Arabia, Turkey, South Africa, Spain, South America, and Newfoundland (Dennison, 1976; Hambrey, 1985; Brenchley et al., 1991; Monod et al., 2003). Diamictites interpreted to be ancient glacial tillite deposits, dropstones in argillites, preserved pavements with glacial scouring and striae, striated pebbles, varved sediments, evidence of global sea level fall, and shifts to the famous Hirnantian cold-water faunas are widespread in these areas (Lenz, 1976; Brenchley and Newall, 1980; Hambrey, 1985; Caputo and Crowell, 1985; Brenchley, 1988; Lavoie and Asselin, 1998; Pope and Steffan, 2003). The major episode began in the Caradocian and lasted through the Ashgillian into the Llandoverian (14-16 my), although the period of most intense glaciation may have only lasted a few million years, peaking in the Hirnantian at 439 Ma (Hambrey, 1985; Brenchley, 1988; Crowell, 1999; Pope and Steffan, 2003). There is evidence for several discrete glacial periods, such as three separate tillites in North Africa (Hambrey, 1985; Brenchley, 1988), with local evidence for sea level oscillation within the main event (Brenchley, 1988). Long (1993) found evidence for at least five phases of shallowing in the Late Ordovician-Early Silurian of Anticosti Island, which he interpreted to represent short-term fluctuations in global sea level directly related to volume changes within the continental ice sheets. Subsequent melting of the continental-scale glaciers caused global glacio-eustatic sea level rise in the earliest Silurian, reflected by widespread rapid transgression and onlap of shallow water carbonates with different faunas (Berry and Boucot, 1973; Lenz, 1976; Brenchley and Newall, 1980).

Causes of Late Ordovician Glaciation

Most authors suggest that the Late Ordovician Gondwanan continental glaciation was likely triggered by the rapid migration of the Gondwanan continent over the South Pole, (Crowell, 1981; Caputo and Crowell, 1985; Hambrey, 1985; Crowell, 1999; Pope and Steffan, 2003; www.scotese.com), encouraging abundant snowy precipitation at high latitudes, stimulating icehouse conditions (Gerhard and Harrison, 2001; Armstrong, 2007), and ended when that continent glided off the Pole to a lower latitude. Between Middle Ordovician and Middle Silurian time, Laurentia, Baltica and part of Gondwana were positioned in equatorial zones (Scotese et al., 1979; Scotese et al., 1985; www.scotese.com) and may have contributed to changes in fundamental ocean-heat circulation and distribution. Sheehan (2001) and Berry and Finney (2001) also discussed the "weathering hypothesis" whereby Late Ordovician Taconian orogeny resulted in increased weathering of silicate terrains and increased clastic deposition, causing a decreased atmospheric CO_2 concentration, and initiation of ice-sheet growth. Hermann and Patzkowsky (2001) indicated that the paucity of extensive vegetation would greatly enhance erosion and burial in the Late Ordovician, although a sudden invasion of the uncolonized land by simple plants (for which there is now evidence: see later section) may, in itself, have led to the decrease in CO_2. Crowell (1999) summarized the causes of Late Ordovician glaciation as a combination of rapid tectonic movement of cratonic blocks over the South Pole, accompanied by continental arrangements favouring strong deep-water flow and upwelling. These may have been modified by extraterrestrial orbital arrangements, low atmospheric CO_2 content, and the general lack of land plants in the continental setting (Crowell, 1999).

Effects of Late Ordovician Glaciation

Globally, biotas were severely affected by these events; indeed, the mass extinction at the end of the Ordovician is one of the five major extinction events of the Phanerozoic, eliminating about 86% of extant species and 60% of marine genera in 2 phases (Sheehan et al., 1997; Finney et al., 1999; Sheehan, 2001; Brenchley et al., 1991, 2003). The worldwide Late Ordovician shallowing of seas and common disconformity was followed by rapid Early Silurian onlap, without discordance, suggesting a eustatic rather than tectonic signature (Lenz, 1976; Brenchley and Newell, 1980). There is extensive evidence for a eustatic sea level fall of 30-100 m in the latest Ordovician, exposing the globally-ubiquitous Ordovician carbonate shelf successions and creating a widespread disconformity in many parts of the world (Berry and Boucot, 1973; Lenz, 1976; Brenchley and Newall, 1980; Brenchley, 1988; Long, 1993; Crowell, 1999). Rapid onset and ending of the glaciation eliminated the time necessary for faunas to adapt to new conditions, and mass extinction ensued (Finney et al., 1999; Sheehan, 2001).

In the Upper Ordovician of Anticosti Island, Long (1993) interpreted the effects of multiple sea level changes to represent eustatic effects and a strong negative shift in $\delta^{13}C$ values at the top to indicate the destruction of the ice caps (Long, 1993). On Manitoulin Island, Kobluk (1984) described two paleokarst surfaces immediately beneath the Ordovician-Silurian boundary which were interpreted to represent a global lowering of sea level. Dennison (1976) pointed to the end-Ordovician unconformity in the Appalachian Basin as evidence of Late Ordovician glacioeustatic sea level drop, and suggested that the thin, but unusually extensive, Queenston Formation redbed mudflats provide additional evidence of the dominance of eustasy at this point in the stratigraphy. It is clear that the Queenston Formation was deposited during a period when sedimentation was strongly influenced by the Late Ordovician glacial epoch.

Loess and Loessite as Glaciation-Related Deposits

Definition and Historical Background

Loess is a widespread, thick, homogeneous deposit of nonstratified windblown silt. It is almost always associated with landscapes recently uncovered by glacial recession, but prior to invasion by a dense vegetative mat. Although the term developed from Quaternary studies, research in the last few decades has alerted geologists to the possibility of more ancient loessite deposits, and of the previously-unsuspected complexities of loess deposits. Loess (or loessite, in the ancient record) is a terrestrial clastic deposit which consists predominantly of well-sorted silt particles, 20-50 μ in diameter, and which occurs as wind-laid sheets, of significant thickness and great lateral extent (Smalley and Vita-Finzi, 1968;

Pye, 1987). It is primarily composed of quartz, feldspar, mica and clay, with calcareous cement, distributed in typically homogeneous, weakly stratified and highly porous deposits (Pettijohn, 1975; Pye, 1987), which are closely associated with glacial icehouse epochs, continental climates and arid settings (Tsoar and Pye, 1987; Smalley et al., 2000; Soreghan et al., 2000).

Although the scientific study of loess has a long history, beginning in Germany in 1820, these deposits were first observed and described in China 2000 years ago as "huang-tu" (meaning "yellow earth") and were attributed there to windstorms (Tungsheng, 1988). In 1824, Karl Caesar von Leonhard published his work on the yellow limey soil around Heidelberg and used the term "löss", which was derived from the local name "loesch" (meaning "loose soil", Tungsheng, 1988) (Smalley, 1975; Pye, 1987; Smalley et al., 2000). Ferdinand von Richthofen, who studied the Chinese Quaternary/modern deposits in great detail, marshalled abundant evidence to support his aeolian theory of deposition (Smalley et al., 2000): a) homogeneity of grain size, composition and structure, b) blanketing geometry over previous topography, c) almost exclusive occurrence of angular quartz grains of silt size, d) absence of stratification, but ubiquitous tendency to vertical cleavage, and e) immense number of terrestrial shells, roots/rhizocretions and mammal bones (Richthofen, 1882). He stated that "Origin from water is perfectly unable to explain [these features]...[the processes must be] those which are founded in the energy of the motions of the atmospheric ocean..."(Richthofen, 1882). Chamberlain (1897) first noted the strong association of loess with the borders of former ice sheets in North America and that the loess graded glacier-ward into till, proving that the deposition of the thick silt blankets was connected to glacial action (Smalley, 1975). Work by Tutkovski in 1900 and Obruchev in 1911 showed by direct observation that loess could be deposited by "foehn" (catabatic) winds blowing off icefields over moraine landscapes in glacial settings, and by dust storms in low-latitude desert areas (Smalley et al., 2000).

Holocene/Pleistocene Loess

Loess covers 5-10% of the Earth's present post-glacial land surface, concentrated in temperate zones and semiarid desert regions (Pécsi, 1980; Pye, 1987; Tungsheng, 1988). Loess thickness, mean and median grain size, and coarse silt content all decrease as distance from the sediment source increases (Olsen and Ruhe, 1980). The experimental work of Bagnold (1941) showed that winds tend to effectively separate mixed surface sediments into a) dune sand (>80 μ), and b) aeolian dust (< 80 μ). Loess consists chiefly of quartz particles in a typically narrow size range with grain diameters of ~10-50 μ (medium to coarse silt), which make up about 80% of the deposit (Smalley and Vita-Finzi, 1968; Smalley and Smalley, 1983), and a median size of 20-40 μ (medium silt) (Pye, 1987; Tsoar and Pye, 1987). Generally, the sediments are moderately to well sorted, with a positive skewness (i.e., tail of finer sizes in the clay to fine silt range) (Pye, 1987). There may also be up to 10% sand-sized material and 20% clay particles (Pye, 1987; Tungsheng, 1988). These grains tend to be equant, angular to subangular (perhaps with edge rounding), have conchoidal fracture and may have adhering clay particles (Pye, 1987; Tungsheng, 1988). The sediments generally have high carbonate and low organic contents, and 40-55% original porosity due to the openwork of silt grains held together by clay bridges (Pye, 1987; Tungsheng, 1988). In addition to carbonate cements, secondary carbonate concretions are very common, either as pedogenic nodules in paleosol horizons or as rhizocretions around rootlets (Pye, 1987).

The silt size of loess deposits are amenable to high sorting pressure and moderate-distance transport at low elevations by aeolian processes (Tungsheng, 1988) where they are easily trapped in thick deposits, generally within a few 10's of km from the sediment source (Tsoar and Pye, 1987). A high, sustained rate of transport likely requires bare, unvegetated, unstable surface material of poorly sorted sediments with a high clay and silt content and a high frequency of strong turbulent winds (Pye, 1987; Muhs et al., 2000). To be recognized as loess, there must be a trapping sink depocentre and the net silt accumulation must exceed the rate of syndepositional weathering and pedogenesis (Tsoar and Pye, 1987). Overwhelmingly, much of the world's Quaternary and Holocene loess occurs in mid-latitudes, which experienced Pleistocene glaciation (Glennie, 1970; Tsoar and Pye, 1987). Smalley (1966) proposed that glacial grinding (and the associated frost fracturing) likely provided most loess material in the world's deposits (Smalley and Smalley, 1983; Tungsheng, 1988). Glacial outwash is particularly rich in the silt size fraction, due to this grinding action (Glennie, 1970; Muhs et al., 2000). Most evidence suggests that the deflation of fine material from newly-exposed and unvegetated braided fluvial plains by strong catabatic winds led to loess accumulation rates up to 3 mm/year in Europe, between Scandinavian ice sheets to the north and Alpine glaciers to the south (Pye, 1987).

In the past, loess deposits were treated as single bodies and therefore overgeneralisation and oversimplification was rampant (Ruhe et al., 1971), but now it is known that loess sedimentation was complex, rates of deposition were variable and deposition was interrupted many times at any site (Ruhe et al., 1971). Typical loess shows signs of weak pedogenesis because of the constant high rate of accumulation-aggradation of silt, but enclosed paleosols represent cessation of aggradation and periods of stronger pedogenesis (Tungsheng, 1988). Many loess sections consist of thick unweathered loess units alternating with thin soil horizons, indicating episodic dust deposition on time scales of tens of thousands of years (Tsoar and Pye, 1987). The presence within a loess of a paleosol indicates a period of relative landscape and climatic stability, although it is really representative of a change in the balance between the ongoing processes of silt accumulation and pedogenesis (Kemp, 2001). Therefore, loess sections can be subdivided based on paleosols, which separate the successions into loess (representing dry, cold periods) and paleosols (representing more humid, warmer periods), with the sharp boundaries between representing rapid climatic changes (Pye, 1987; Tungsheng, 1988; Kemp, 2001, Wang et al., 2003). Thus, the thick successions comprise multicyclic paleoclimatic archives (Tungsheng, 1988).

Ancient Loessite

Rautman (1975) made the first identification of ancient loessite in the Lak Member of the Sundance Formation of Wyoming, describing 25 m of Upper Jurassic non-stratified, non-fossiliferous, maroon siltstone and sandy siltstone positioned above a shoreline complex in a regressive succession. He suggested that the a) stratigraphic position and lack of fauna suggested a nonmarine setting, b) well sorted fine texture suggested aeolian suspension transport, c) lack of stratification indicated that the usual tractive processes were not in play, d) lack of channel sandstone bodies with cross bedding indicated no fluvial processes, and e) lack of paleosols meant rapid deposition with little time for pedogenic processes to operate (Rautman, 1975). Edwards (1979) followed with similar arguments for massive to faintly stratified siltstone units intercalated within known Late Precambrian tillites and glaciolacustrine mudstones in Norway. The siltstones were reddish, well sorted, with angular grains of coarse silt to very fine sand size and had a few horizons of carbonate "pebbles" (Edwards, 1979), which may have been calcrete nodules. Similarly, Johnson (1989) provided the most comprehensive description and interpretation of very thick loessite in the Carboniferous-Permian Maroon Formation of Colorado, an area positioned in a low-latitude arid climatic belt during the late Paleozoic glacial epoch. Structureless, homogeneous, well sorted and well-cemented sandy siltstone, primarily of coarse silt grain size were coupled with roots, weakly-developed paleosols, desiccation cracks and raindrop impressions (Johnson, 1989).

During the last 20 years, there have been further studies of Carboniferous-Permian loessite units of equatorial Pangea, exposed in the western US, which were coeval with high latitude Pangean glaciation. These laterally continuous sheets of remarkably restricted grain size distribution, dominated by medium to coarse silt, correlate well with aeolian sandstone units positioned upwind and interpreted to represent windy, cold, arid phases of glacial climatic cycles (Soreghan, 1992; Kessler et al., 2001). They are separated by rooted paleosols which marked cessation of silt influx, coupled with landscape stability and revegetation, attributed to warmer, wetter interglacial phases (Beck et al., 2000; Kessler et al., 2001). The extensive, uniform, massive unstratified red siltstone of the Triassic Mahogany Member, Ankareh Formation of Utah was interpreted by Chan (1999, 2000) as a low-latitude Pangean loessite capped by an immature paleosol. Moderately sorted, angular to subangular, medium siltstone to very fine sandstone is primarily quartz in the medium to coarse silt range, with abundant hematite matrix and carbonate cement (Chan, 1999). These deposits were interpreted to represent deflation of nearby playa sediments during an arid windy period, and trapping in a localized subsiding basin (Chan, 1999).

Many of the aforementioned features are characteristic of the Milton member, suggesting the alternative interpretation of these deposits as another example of ancient loessite.

Ordovician Paleosols: Identification and Interpretation

Pedogenetic Processes

Soil is the surface material altered by physical and chemical weathering and by biogenic action, which result in disaggregation of the original surface sediments (Retallack, 1981). The presence of a paleosol represents a phase of landscape stability with neither significant deposition nor erosion (INQUA, 1990). Many paleosols are simple horizons separated by non-pedogenic deposits resulting from short-lived phases of deposition separated by long periods of no net sedimentation, during which pedogenesis occurred (Allen and Wright, 1989). Reddening ("rubification") is common in soils of arid to semi-arid climates, typically with a diffuse lower boundary and an upward increase in colouration (Retallack, 1981; Allen and Wright, 1989). Crusted, nodular or concretionary concentrations of carbonate, iron and silica are very common in certain soil profiles (Van Houten, 1982). Irregularly-shaped, hard glaebules of pebble-sized micrite (calcrete nodules) are important in arid climate soils (Esteban and Klappa, 1983; Retallack, 1988). Horizons of carbonate concretions, which are common in nonmarine redbed successions such as the Milton member, develop through stages from an immature horizon of dispersed carbonate nodules to a mature bed of indurated, impermeable carbonate.

In the early stages of soil development, pedoturbation processes begin to destroy the original layering of the sediment or parent material (Allen and Wright, 1989), generally within a few hundreds to a few thousand years and the resulting destratified hackly, massive to blocky jointed texture of the pedogenic sediment is quite distinctive (Retallack, 1981; Van Houten, 1982; Retallack, 1988). Discrete elongate columnar and prismatic aggregates, called "peds", can result from repeated opening and closing of an array of desiccation cracks of various dimensions. The peds are defined by an irregular network of planes, called cutans, which are typically lined by thin skins of clay or minerals marked by randomly-oriented slickensides (Retallack, 1988, 1992) (Figure 11). Definable peds, cutans and slickensides, observed in the Milton member as described herein, are typical of periods of intense pedogenesis in seasonally-wetted soils (Retallack, 1988).

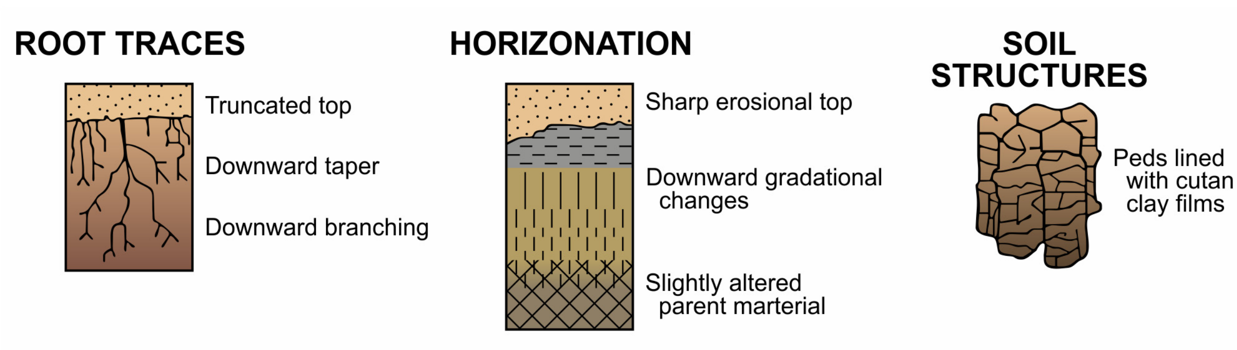

Figure 11. Important diagnostic paleosol features (modified from Retallack, 1992).

Several biogenic features are considered diagnostic of subaerial exposure in the terrestrial environment. Rootlets, or root molds/rhizocretions and the presence of a dark organic-rich upper layer are key diagnostic criteria in post-Cambrian deposits (Retallack, 1981; Van Houten, 1982; Esteban and Klappa, 1983). The lack of definable structures or deeply-penetrating rhizocretions are more typical of seasonally-wetted/ better drained areas, and reduction haloes around unpreserved, former root traces are common in arid climate red paleosols (Retallack, 1988).

Classification and Utility of Paleosols

In a geological time frame, paleosols form quite rapidly (hundreds to thousands of years) and therefore can be ideal correlation and sequence stratigraphic tools in continental sequences (Kraus, 1999). Mack et al. (1993) proposed a classification system, primarily for ancient soils, which relies on the presence of morphological properties, textures and stable minerals likely to be preserved over time. The nine orders of paleosol (Mack et al., 1993) include the following three, which may be relevant to study of the Milton member of southern Ontario. (i) *Protosols* are paleosols with weak development of horizons related to homogenization by pedoturbation (could be very common, but difficult to identify positively due to lack of distinctive characteristics), (ii) *Vertisols* are paleosols whose most prominent characteristic is lack of horizonation due to homogenization of the profile by pedoturbation and shrink/swell dynamics of abundant expandable clays (characterized by dominantly vertic features such as desiccation cracks, wedge-shaped aggregates or peds, slickensides and cutans, all of which form quite rapidly during initial destratification due to the high clay content, and seasonal wet/dry cycles (Duchafour, 1982; Buol et al., 1989; Allen and Wright, 1989)). (iii) *Calcisols* are paleosols which are characterized by significant accumulations of soluble minerals as discrete horizons, precipitated in the vadose zone (occurring as discontinuous veins, scattered nodules, or massive nodular beds, representing the early cementation of soil matter by carbonate (sometimes gypsum) and are widespread in warm, arid to semi-arid climates (Retallack, 1981; Van Houten, 1982; Esteban and Klappa, 1983; Allen, 1986)).

Late Ordovician Evolution of Early Land Plants, Soils and Terrestrial Ecosystems

Soils and biological ecosystems have evolved in tandem through the past 450 million years of geological time. Higher plants could not colonize the continents until they developed 1) surficial waxy cuticle to deal with desiccation, 2) sporopollenin to provide dessication-resistant spores, 3) rigid lignin to provide structural strength and vascular tracheids, 4) the ability to synthesize flavinoids and phenylpropanoids to absorb UV damage, and possibly, 5) a symbiotic relationship with fungi to acquire nutrients from barren rock rather than surrounding water (Knoll et al., 1986). The first three of these characteristics created material which could be preserved in the hostile terrestrial environment and identified in the fossil record (Wellman, 2003). The first group of plants to evolve at least some of these features were the Bryophytes (one of the three great divisions of the Plant Kingdom and among the simplest of terrestrial plants), including mosses, hornworts and liverworts. These most likely emerged from cyanobacteria or charophytic green algae and crept onto the land in the mid-Ordovician (Taylor, 1981; Knoll et al., 1986; Retallack, 1986; Wellman, 2003). The widespread appearance of small, obligate tetrad spores and fossil cuticular material in rocks of Llanvirnian-Caradocian-Ashgillian ages attests to the rapid colonization of these founder organisms (Gray, 1985; Knoll et al., 1986; Gray, 1993, Wellman, 2003). This crucial emergence of the terrestrial bryophytes had a profound effect on the development of Late Ordovician landscapes and soils and was one of the most important events in the history of life on Earth, initiating the development of complex terrestrial ecosystems (Wellman, 2003). This revolution was occurring during deposition of the Queenston Formation. The first humble bryophytic plants, stabilizing and transforming the land and terrestrial environments forever, quickly led to more prolonged periods of pedogenesis and more distinct structured paleosols (Retallack, 1985, 1986; Gray, 1993). Pre-

Silurian soils are difficult to recognize because of the lack of vascular plants with roots which could stabilize, bioturbate and chemically-alter the substrate (Feakes and Retallack, 1988). They also show evidence of some of the oldest terrestrial trace fossils in the world (Late Ordovician Juniata Formation in Pennsylvania, coeval to the Queenston Formation) (Retallack and Feakes, 1987), excavated by bilaterally symmetrical, millipede-like organisms (Retallack and Feakes, 1987). Modern millipedes still thrive on non-vascular plants where nutritional value and competition by insects is limited – both liverworts and millipedes still succeed in the modern world "because they were there first and they are admirably suited to each other" (Retallack, 2001).

Late Ordovician Paleosols of Eastern North America

The Late Ordovician Juniata Formation of the northeastern U.S. (a more proximal correlative of the Queenston Formation) has yielded excellent evidence of simple paleosols, simple animal burrows and traces of the first land vegetation. Feakes and Retallack (1988) and Driese and Foreman (1992) discussed the occurrence of weakly-developed Calcisols and Vertisols in the Juniata Formation, formed in a subhumid to semi-arid climate and a low latitude setting. The Juniata represents a fluvial-alluvial redbed floodplain, extending from the newly-developed high Taconian orogen in the east to a marine shoreline to the northwest in southern Ontario (as the Queenston Formation) (Retallack, 1985; Feakes and Retallack, 1988). These soils are thin and dark reddish brown, iron-rich and highly oxidised, with vertic features and a hackly appearance (Retallack, 1985; Driese et al., 2001). Ferruginized fecal pellets and abundant liverwort spores have been recovered, and light carbon and oxygen isotopes indicate a terrestrial setting with high soil respiration, suggesting large populations of animals and heterotrophic microbes (Retallack, 2001). Retallack (2001) discovered *Scoyenia* burrows in the Juniata, whereas Driese and Foreman (1991) identified *Skolithos*- and *Glossifungites*-like trace fossils in coeval deposits to the south. The Late Ordovician Queenston Formation of southern Ontario includes pervasive development of vertisols and some calcisols in brick red siltstones, as described here.

The loessite concept and the Milton member: a new interpretation

Loessites can be expected to be associated with all glacial icehouse epochs, including Late Precambrian, Late Ordovician and Carboniferous-Permian (Johnson, 1989; Soreghan et al., 2000), both in paraglacial and low-latitude arid settings. These unique aeolian deposits should be best preserved when occurring stratigraphically above a shoreline tract, and beneath a regional marine transgression or regional unconformity (Kocurek, 2000). Possible detailed sedimentological characteristics to look for in ancient strata, on regional, local and sediment body scales (as described in previous pages), and their comparison to known characteristics of the upper Queenston Formation, Milton member are included in Table 1. The Milton member displays many of these expected characteristics, and is here interpreted to represent a glacial-associated loessite deposit.

The Queenston Formation is of Late Ordovician age, a known glacial epoch, and was deposited when landscapes were largely barren of thick vegetation, in a low-latitude setting with an arid to semi-arid climate, positioned on the downwind side of the Taconian highlands relative to prevailing Southeast Trade Winds. Those highland sediment source areas included abundant poorly sorted molasse and fluvial/alluvial sediments, and may have housed alpine glaciers and their deposits. Regionally, the deposits in question represent the distal equivalents of known fluvial/alluvial Juniata Formation deposits, overlie a regressive sequence of shallow marine to shoreline facies and, in turn, are overlain by a regional subaerial unconformity and fluvial facies of the Whirlpool Formation. In gross geometry, the Milton member comprises a large tabular blanket geometry, of modest thickness, but extending laterally for 300 km in all directions, overlying a regressive succession and overlain by a subaerial unconformity. It is generally very uniform, massive, nonstratified and homogeneous on a regional scale, with no significant fluvial channel sandstones known anywhere. The detailed sedimentology is also unique. The deposits are dominated by angular to subangular medium to coarse grained silt (a hallmark of loess and loessite), particularly of dolomite and quartz mineralogy, set in a hematitic clay matrix. Weakly-developed calcisols, calcrete horizons, vertisols, dominance of vertic features, possible rhizocretions and rare aeolian ripples are present, all suggesting subaerial deposition. However, no channel sandstone bodies with cross bedding are evident, suggesting fluvial processes were not important. I believe these characteristics suggest the Milton member may represent an ancient loessite deposit.

If this interpretation is correct, the Milton member of the Queenston Formation represents the first ancient loessite identified in Canada, one of the oldest loessites in the world, and the first anywhere to be associated with the Late Ordovician glacial epoch. From this conclusion, further questions arise. Late Ordovician continental glaciation caused global sea level lowstand, exposing naked unvegetated cratons and sedimentary cover at their margins to pervasive erosion and aeolian deposition. Were pre-Middle Silurian landscapes pre-configured for loess generation, transport and deposition due to the limited presence and primitiveness of the Late Ordovician vegetation (which resided primarily in coastal areas)? Similarly, did deposition of Late Ordovician glacial loess pre-configure land areas for the great plant invasion of the Late Ordovician-Middle Silurian due to the unusual fertility of these same loess soils? Was there an iterative feedback mechanism between these two factors which intimately connected them? Coincidentally, did this intercontinental transport of glacial-loess dust (and any entrained spores) across great distances, itself help to seed the barren Late Ordovician landscapes?

TABLE 1. CHARACTERISTICS OF ANCIENT LOESSITE DEPOSITS COMPARED TO THOSE OF THE MILTON MEMBER.

Ancient Loessites	Milton member
Overall association with terrestrial continental and marginal marine settings, and regional subaerial unconformity	Overlies regressive sequence of shallow marine to shoreline deposits, is overlain by regional subaerial unconformity and fluvial Whirlpool Formation, is distal equivalent of fluvial/alluvial Juniata Formation
Association with pre-Middle Silurian nonmarine deposits due to limited primitive terrestrial vegetation, which was generally present only in coastal areas	Pre-Middle Silurian (Upper Ordovician), landscapes were apparently fairly barren
Association with glacial epochs	Late Ordovician is known glacial icehouse epoch
Arid to semi-arid and continental paleoclimates	Arid to semi-arid paleoclimate
Associated with glacial deposits in mid latitudes, or desert, redbed, evaporite and fluvial/alluvial deposits in low latitudes	Position in low latitudes; is a clastic redbed sequence, associated with fluvial/alluvial deposits, in a predominantly carbonate succession with minor evaporite occurrences
Paleogeographic position down paleo-wind of major silt source such as desert dunes, mixed fluvial deposits, highland area, or known glacial deposits	Located downwind (relative to prevailing Southeast Trade Winds) of Taconian molasse shed from rising Taconian highlands, may have been alpine glacial deposits present
Tabular blanket, laterally continuous geometry, with thickness decreasing away from the sediment source	Large tabular blanket geometry, laterally continuous, thickness decreases to NW away from sediment source
Relatively thick, but much more widespread than modest thickness would initially suggest	Modest thickness of several tens of metres, but extending laterally for 300 km in all directions
Typically uniform and homogeneous on regional scale, massive and structureless, thick bedded	Very uniform and homogeneous on regional scale, generally massive, thick- or unbedded
Lack of direct presence of dominant fluvial channel deposits within sediment body	No fluvial channel sandstones of any thickness known anywhere
Paleogeographic position at location of aeolian sediment trap mechanism such as topographic obstacle, subsiding basin, moist coastal ground, or vegetated surfaces	Paleogeographic position between highland-derived fluvial/alluvial plain and marine shoreline/sabkha plain
Very restricted and well sorted grain size, in medium to coarse silt grade (10-50 μ), equant/bladed grains are angular to subangular (may be rounded sand grains)	Restricted grain size, dominated by silt in 10-100 μ range, equant/bladed grains are angular to subangular
Mineralogy dominated by quartz, followed by feldspar, possible abundance of mica, usually very carbonate-rich and organic-poor, often a hematite matrix	Mineralogy dominated by dolomite grains, followed by quartz, set in hematitic clay matrix, very carbonate-rich
Very porous and permeable in its original state, but may be very clay-rich if heavily altered by diagenesis	Very clay-rich, but likely diagenetic
Paleosols very common , particularly weakly developed inceptisols, ubiquitous vertic features, calcisols may be present at specific horizons which mark individual depositional phases	Common weakly developed calcisols and horizons of calcrete nodules, plus ubiquitous vertic pedogenic features
Roots, insect and arthropod burrows, mammal bones, and subaerial trackways are common, especially in the paleosols, and low-amplitude/long-wavelength aeolian ripples may be present at particular horizons	Deposit pre-dates many biological indicators, but possible rootlets and/or rhizocretions, small rare burrows, evaporitic crystal molds, few rare aeolian ripples

Integrated Regional Interpretation of the Queenston Formation

The red bioturbated mudstone (Facies 1), typical of the lower Queenston (Streetsville member), is interpreted to reflect the background, low-energy slow deposition of silts and muds in a marine oxidizing environment where burrowing was extensive. In many ways this deposition was a continuation of that typical of the underlying Georgian Bay Formation environment. The red colouration is assumed to be post-depositional diagenetic in nature. The reddish fine sandstone beds (Facies 2), which are ubiquitous throughout the lower and middle Queenston (Streetsville and Bronte Creek members) but uncommon in the upper Queenston (Milton member), occur in thin, sharp-based beds with evidence of deposition from traction currents flowing toward the WNW, and are interpreted as representing rapidly-emplaced, higher energy density current deposits in the shallow marine setting, likely sourced from the Taconian orogen in the southeast. The presence of HCS at the tops of sequences and toward the middle of the formation suggests these currents may have been related to storm deposition, and that some deposition took place in the shoreface to shoreline setting. Thin, bioclastic calcarenites (Facies 3), common in the lower Queenston and in the northwest region (Streetsville and Bronte Creek members), are also generally interpreted as representing rapidly-emplaced, higher energy density current deposits in the shallow marine setting. These may have been sourced from a different, carbonate-rich area of the shelf.

Thick developments of brick-red pedogenic medium to coarse grained siltstone (Facies 4), so characteristic of the Milton member of the upper Queenston, are somewhat enigmatic, but here I have presented a novel interpretation. The characteristic blocky, rubbly to massive, commonly vertically-fractured macro-texture is here interpreted to represent the results of multiple pedogenic soil-forming processes in a subaerial setting, as protosols or vertisols. Peds and cutans, which are hallmarks of the soil-forming processes, are well-developed. Small, rounded, greenish-white, calcareous nodules and glaebules, interpreted as caliche nodules, are commonly scattered throughout units of red mudstone, or may appear as discrete calcrete horizons, or calcisols. The lack of bedding or lamination and of significant bioturbation, and the dominance of vertic features, is consistent with dominance of post-depositional pedogenic processes. Thin siltstone to sandstone beds typically have been altered by the same pedogenic processes affecting the background siltstones. There is a distinct lack of evidence of significant fluvial systems or input associated with these thick siltstone deposits. There are even hints of Upper Ordovician roots, although this would be very unusual. These deposits are here interpreted to represent the thick development and preservation of pedogenically-altered loess in a glacially-influenced, subaerial environment.

The sediments of the Streetsville and Bronte Creek members are interpreted as having been deposited in an overall shallowing marine-to-nonmarine setting, as suggested by the typical thickening- and coarsening-upward trend of interbedded mudstone and siltstone. Individual thickening- and coarsening-upward sequences suggest this occurred through a series of lesser successive shallowing events. These shallow marine to shoreline sediments were overlain by the thick finer grained, subaerial loess deposits of the Milton member. At the top of the Queenston, the regional Cherokee Unconformity represents the subaerial peneplanation of the Tippecanoe I Foreland succession, which was later sharply overlain by the Whirlpool and Manitoulin formations during marine transgression of the Tippecanoe II Foreland succession.

CONCLUSIONS

1. The Queenston Formation comprises red to greyish-red, non-calcareous mudstone to siltstone with thinly interbedded red fine grained sandstone and minor bioclastic limestone, arranged in stacked thickening- and coarsening-upward successions in the lower two-thirds, but dominated by red uniform siltstone in the upper one-third. Facies included are 1) red, bioturbated shallow marine mudstone, 2) green to red very fine to fine grained sandstone deposited in shallow marine and shoreface settings from traction density currents, 3) bioclastic, shallow marine limestone deposited from traction density currents, and 4) uniform red, pedogenically-altered siltstone (Milton member), here interpreted for the first time to represent deposition as thick units of loess in a glacially-influenced subaerial coastal plain setting.
2. New paleocurrent data (including 105 direct paleoflow indicators and 70 indirect trend indicators) suggest a regional shoreline trend of 20°/200° with an offshore regional paleoslope direction of 310°.
3. Important new sedimentological concepts applicable to the Queenston Formation (Milton member) emerged from this study. The new interpretive notions of paleosols and of loessite, not previously applied to the Queenston of Ontario, proved to be crucial to the interpretation of the upper portion of this Formation, which was deposited in a subaerial setting immediately prior to a major regional subaerial unconformity and during a time of glacially-influenced global climate. Paleosol types include pervasive vertisol development and numerous discontinuous calcisol horizons. Loessite of the Milton member is manifest as a tabular blanket of uniform, nearly massive or unbedded siltstone with peds, cutans and paleosols, which is located stratigraphically above regressive shoreline facies and beneath a regional subaerial unconformity at a time period of global glaciation.
4. The abundant calcisols and strong rubrification in the upper Queenston imply a generally arid-semi-arid climatic regime, located in a low-latitude setting, in a rain-shadow area.

ACKNOWLEDGEMENTS

I wish to thank Peter Telford for introducing me to the wonders of working on Niagara Escarpment outcrops many years ago during our time at Ontario Geological Survey, and to thank GSC for allowing me to later pursue these strata with additional fieldwork. Terry Carter introduced me to the excellent geological archives contained in the subsurface cores stored at the Ontario Oil, Gas and Salt Resources Library in London. Dave Sargent expertly created the figures, improved the overall presentation and handled assembly and production of the Open File. Rob MacNaughton provided an incisive review of this study which helped improve the manuscript. Many thanks to all of you.

APPENDIX I. QUEENSTON FORMATION CORES (10)

(Core Name, Queenston Fm Thickness, Lat./Long, 1:50000 NTS Sheet (NAD 83), Grid/Block Location)

1. Corbetton #1 (Shelburne), 97 m, 44°09'N / 80°21'W, 41A/1 Dundalk, 517877
2. OGS 83-1 (Milton), 149 m, 43°32'N / 79°58'W, 30M/12 Brampton, 838212
3. OGS 82-1 (Courtright), 75 m, 42°45'N / 82°30'W, 40J/16 Sarnia
4. OGS 82-2 (Chatham), 94 m, 43°20'N / 82°05'W, 40J/8 Chatham
5. OGS 82-3 (Port Stanley), 137 m, 42°40'N / 81°10'W, 40L/11, Port Stanley
6. OGS 82-4 (Wiarton), 22 m, 44°50'N / 81°13'W, 41A/14 Cape Croker, 829578
7. US Steel #1 (Port Dover), 178 m, 42°40'N / 80°20'W, 40I/9 Long Point
8. Richardson #522-25 (Wiarton), 53 m, 44°53'N / 81°10'W, 41A/14 Cape Croker, 865698
9. Anschutz Lake Erie 6-R, 1 m at top of Queenston, 42°50'N / 79°25'W, 30L/14, Block 6-Tract R-Qtr 4
10. Anschutz Lake Erie 22-S-4, 2.5 m at top of Queenston, 42°45'N / 79°30'W, 30L/13, Block 22-Tract SQtr 4

APPENDIX II. QUEENSTON FORMATION OUTCROPS (53)

(Outcrop Name, Queenston Fm Thickness, Lat./Long, 1:50000 NTS Sheet (NAD 83), Grid Location)

1. Niagara River, Queenston, 20 m, 43°10'N / 79°03'W, 30 M/3/6 Niagara, 587808
2. Sir Adam Beck Power Station Roadcut, 2 m, 43°08'N / 79°03'W, 30 M/3/6 Niagara, 581776 – 591785
3. Ball's Falls, 4.5 m, 43°08'N / 79°23'W, 30 M/3/6 Niagara, 317771
4. Niagara River, The Whirlpool, 1.5 m, 43°07'N / 79°04'W, 30 M/3/6 Niagara, 571763
5. Highway 403 Roadcut, 1 m, 43°15'N / 79°56'W, 30M/4 Hamilton-Grimsby, 865886
6. Jolley Cut, 1 m, 43°15'N / 79°52'W, 30M/4 Hamilton-Grimsby, 926887
7. Red Hill Creek, 10 m, 43°14'N / 79°47'W, 30M/4 Hamilton-Grimsby, 987875
8. Devil's Punchbowl, 14 m, 43°13'N / 79°46'W, 30M/4 Hamilton-Grimsby, 850010
9. Kenilworth Rd, 12.5 m, 43°14'N / 79°50'W, 30 M/4 Hamilton-Grimsby, 952875
10. Sixteen Mile Creek, Glenorchy, 7 m, 43°28'N / 79°48'W, 30M/5 Hamilton-Burlington, 970142
11. Sixteen Mile Creek Roadcut, Glenorchy, 13 m, 43°29'N / 79°46'W, 30M/5 Hamilton-Burlington, 991152
12. Sixteen Valley Conservation Area, 9 m, 43°30'N / 79°40'W, 30M/5 Hamilton-Burlington, 990171

13. Bronte Creek, Zimmerman Bridge, 6 m, 43°30'N / 79°51'W, 30M/5 Hamilton-Burlington, 935090

14. Bronte Creek Quarry, 11 m, 43°25'N / 79°47'W, 30M/5 Hamilton-Burlington, 969078

15. Bronte Creek Provincial Park, 33 m, 43°25'N / 79°46'W, 30M/5 Hamilton-Burlington, 993073

16. Sixteen Mile Creek, Lion's Valley Park, 20 m, 43°28'N / 79°45'W, 30M/5 Hamilton-Burlington, 010126

17. Sixteen Mile Creek, Golf Club 7 m, 43°27'N / 79°42'W, 30M/5 Hamilton-Burlington, 050119

18. Sixteen Mile Creek, Sunningdale, 29 m, 43°28'N / 79°43'W, 30M/5 Hamilton-Burlington, 035125

19. Sixteen Mile Creek, Dorval Rd., 11 m, 43°27'N / 79°42'W, 30M/5 Hamilton-Burlington, 046117

20. Bronte Creek, QEW, 16 m, 43°21'N / 79°44'W, 30M/5 Hamilton-Burlington, 021069

21. King Road Quarry, Burlington, 10 m, 43°20'N / 79°50'W, 30M/5 Hamilton-Burlington, 932989

22. Grindstone Falls, Waterdown, 7 m, 43°19'N / 79°53'W, 30M/5 Hamilton-Burlington, 902979

23. Aldershot Quarry, 20 m, 43°20'N / 79°51'W, 30M/5 Hamilton-Burlington, 929977

24. Credit R., W. of Huttonville, 4.8 m, 43°39'N / 79°49'W, 30 M/12 Brampton, 953332

25. Streetsville Quarry, 15 m, 43°35'N / 79°44'W, 30 M/12 Brampton, 022261

26. Credit River, Georgetown RR, 8 m, 43°40'N / 79°53'W, 30 M/12 Brampton, 892345

27. Credit River, Georgetown Dynamo, 9 m, 43°40'N / 79°54'W, 30 M/12 Brampton, 881349

28. Terra Cotta Roadcut, 2 m, 43°44'N / 79°57'W, 30 M/12 Brampton, 851428

29. Terra Cotta Badlands, 13 m, 43°42'N / 79°57'W, 30 M/12 Brampton, 842397

30. Hilltop Quarry, 0.5 m, 43°41'N / 79°57'W, 30 M/12 Brampton, 843384

31. Milton Heights Quarry, 12 m, 43°31'N / 79°55'W, 30 M/12 Brampton, 872180

32. Credit River, Norval, 8 m, 43°39'N / 79°51'W, 30 M/12 Brampton, 930335

33. Red Cliffs, 9.5 m, 43°39'N / 80°52'W, 30 M/12 Brampton, 905343

34. Cheltenham Quarry, 8 m, 43°44'N / 79°55'W, 30 M/12 Brampton, 858435

35. Queenston Badlands, Inglewood, 20 m, 43°46'N / 79°57'W, 30 M/13 Bolton, 849472

36. Hockley Roadcut, 14 m, 43°58'N / 80°05'W, 40 P/16 Orangeville, 742681

37. Glen Cross, 4 m, 43°59'N / 80°03' W, 40 P/16 Orangeville, 765705

38. Cataract, 8 m, 43°49'N / 80°02'W, 40 P/16 Orangeville, 785542

39. Mono Centre Roadcut, 6 m, 44°02'N / 80°04'W, 41 A/1 Dundalk, 753765

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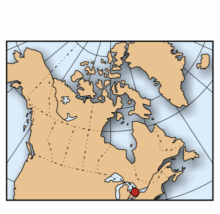
UPPER ORDOVICIAN/LOWER SILURIAN – SOUTHERN ONTARIO
COLLINGWOOD/BLUE MOUNTAIN/GEORGIAN BAY/QUEENSTON/MANITOULIN/CABOT HEAD/FOSSIL HILL/ROCHESTER FMS
OGS Port Stanley 82-3 (1983)
Lot 9, conc. I, Yarmouth Twp., Elgin Co.
Lat. 42°40'N Long. 81°10'W

Figure 5. Typical core measured section of Queenston Formation, OGS 82-3 Port Stanley core.



doi:10.4095/298817

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2016

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Recommended citation
Hamblin, A.P., 2016. A Late Ordovician ancient loessite deposit in a glacially-influenced setting?: the Milton member of the Queenston Formation in southern Ontario, and synthesis of the background concepts behind a novel interpretation; Geological Survey of Canada, Open File 7860, 42 p. + figure. doi:10.4095/298817

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