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Stratigraphy of the Upper Silurian to Middle Devonian, Southwestern Ontario

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Abstract

The upper Silurian–Middle Devonian succession was dominated by carbonate and evaporite deposits, with minor siliciclastic sedimentation, and a significant hiatus across the Silurian-Devonian (S-D) boundary in southwestern Ontario. The stratigraphic units include, in ascending order: Late Silurian Bass Islands/Bertie formations and Salina G Unit, the Devonian Oriskany Formation, Bois Blanc Formation (including Springvale Member), Detroit River Group (including the Lucas, Amherstburg and Sylvania formations), Onondaga Formation, and Dundee Formation.

Below the S-D unconformity, the upper Silurian Bass Islands/Bertie formations are predominantly dolostone of peritidal-sabkha origin and episodic subaerial exposure. Revised stratigraphic correlation shows that the Bertie Formation is older than the Bass Islands Formation, and the regional uplift of the study area has eroded the top of the Silurian deposits in the Niagara Peninsula. Four types of paleokarst systems are recognized in the strata below the S-D unconformity: 1) evaporite-dissolution; 2) syn-depositional brecciation; 3) surface karstification below S-D unconformity; and 4) deep burial dissolution. Climatic conditions during S-D boundary interval were probably semi-arid and the regional uplift of the North American craton led to marine regression and karstification.

The S-D compound boundary comprises three unconformity surfaces. In the Niagara region, these unconformities are marked by three levels of regionally distributed siliciclastic sandstones, including the Oriskany Formation, Springvale Member of the Bois Blanc Formation and an unnamed sandstone unit below the Onondaga Formation. These units are interpreted as eolian sandstones, reworked and remobilized during the regional-scale regressions and exposure of large parts of Laurentia. In the Michigan Basin, the lower two unconformities merge into one and the upper sandstone unit may be coeval with the Sylvania Formation.

In Ontario, the Lower-Middle Devonian Onondaga Formation is restricted to the Niagara area. Onondaga facies reveal fossiliferous carbonate deposition on a carbonate ramp sloped easterly into the Appalachian Basin centre. Temporally, the Onondaga facies in Ontario reflect one third-order sequence. The open marine facies of the Onondaga Formation grade southwestward into the intracratonic and lagoonal facies of the Amherstburg Formation and the Lucas Formation. The Dundee Formation in Niagara area can be correlated to the upper Moorehouse and Seneca Members of the Onondaga Formation, evidenced by the conodont data and the regional occurrence of Tioga
ash bed. The sequence stratigraphic framework shows that the depositional systems are influenced by both eustatic changes and the migration of paleo-highlands in Ontario.

**Key Words:** carbonate and evaporite, paleokarst, Silurian-Devonian boundary, stratigraphy, southwestern Ontario
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CHAPTER 1. Introduction

1.1 Introduction and General Purpose

Uppermost Silurian-Middle Devonian strata underlie much of southwestern (SW) Ontario (Figure 1.1). They are dominated by carbonate rocks (limestone and dolostone) with evaporitic and siliciclastic intervals (Johnson et al. 1992). Various deep bedrock potable groundwater flow zones and formational fluids across SW Ontario are developed within these strata, including some of the deepest fresh or potable bedrock groundwater flow zones (Brunton and Dodge 2008; Carter and Brunton 2011; Carter and Fortner 2012). The Lower-Middle Devonian carbonate strata are also important oil-producing intervals in SW Ontario (Bailey Geological Services Ltd. and Cochrane 1985). Such sedimentary successions with several major unconformities provide a basis for the reconstruction of regional paleogeography, paleoecology and paleoclimate, as well as the tectonic evolution of North America in Late Paleozoic. Therefore, the Ontario Geological Survey (OGS) initiated a multi-year bedrock groundwater mapping program on these strata from 2013 to 2019, aiming at improving the geological understanding of subsurface bedrock aquifers, raising residents’ awareness of potable water protection, and enhancing the quality of public health service.

This thesis project is partly integrated into the aquifer mapping program, supported by a three-year regional deep bedrock drilling program across SW Ontario starting from 2018-2020. This study comprises petrography, paleo-environment interpretation and sequence stratigraphy of the uppermost Silurian-Middle Devonian strata in SW Ontario to refine the sequence stratigraphic relationships of various sedimentary packages and to delineate geological controls on those subsurface bedrock aquifers. This study incorporates data from 38 continuously cored holes (see Sun et al. 2014, 2015, 2016, 2017; Table 1.1) and hundreds of bedrock and oil/gas well records across SW Ontario to provide the comprehensive sedimentological and stratigraphic perspectives
of the study area. The stratigraphic units that have been studied in detail include, in ascending order, the Silurian Salina G Unit and Bass Islands/Bertie formations, the Devonian Oriskany Formation, Bois Blanc Formation (including its lower Springvale Member), Detroit River Group (including the Lucas, Amherstburg and Sylvania formations) and its counterpart Onondaga Formation in Niagara Peninsula, and Dundee Formation (Fig 1.1; see Armstrong and Carter 2010). The study area straddles two basins, the Michigan Basin to the west and the Appalachian Foreland Basin to the east. It is geographically outlined by three sides by Lake Huron, Lake Erie and Lake Ontario, respectively, and its eastern side is bounded on the Niagara Peninsula adjacent to New York (NY) State (Fig 1.2 and 1.3).

1.2 Problems and Objectives

Despite about 150 years of study, the understanding of the relative ages, lithofacies distribution and correlation of the Michigan stratigraphic units into the Appalachian units remains challenging for geologists, sedimentologists and petroleum explorationists. The progress has been hampered by a number of factors: 1) there are limited number of existing cored holes and lack of good outcrops, so there have been virtually no investigations of regional geological controls on bedrock groundwater flow zones that extend from Middle Devonian carbonates to Salina G horizon; 2) diagenetic overprinting of post-depositional processes related to basinal fluid flow and mixing with freshwaters, such as karst-related erosion, chertification and dolomitization, inhibits the observation; 3) a unified stratigraphic nomenclature is lacking between subsurface and surface studies and across geopolitical boundaries (Canada and the USA); 4) biostratigraphy in the uppermost Silurian–Middle Devonian evaporite and carbonate successions is poorly constrained (see Uyeno et al. 1982); and 5) correlations are difficult because of the lithology changes rapidly, both laterally and vertically, on the far-field side of the Appalachian Foreland Basin, which are
controlled by intermittent forebulge migration and relative sea level changes.

To tackle these problems, the fundamental goal of this thesis is to refine regional subsurface architecture and stratigraphy of the upper Silurian through to Middle Devonian stratal units that are related to various bedrock potable groundwater flow zones and formational fluids across SW Ontario. Although the thesis will not deal directly with groundwater mapping, the main purpose is to provide a regional sequence stratigraphic framework for these studies. The main objectives are: (1) to describe lithofacies units in both boreholes and outcrops, and to correlate the lithofacies within a carbonate-mixed siliciclastic sequence stratigraphic framework to better understand their distribution, depositional environments and the relationship to relative sea-level changes; (2) to refine regional stratigraphic correlation by tracing the stratigraphic units from the outcrop belt in the Niagara Peninsula and western New York State into the subsurface of SW Ontario across the edges of both the Appalachian Foreland Basin and the Michigan Basin; (3) to summarize the lithostratigraphy and biostratigraphy incorporated with the works done in New York, Ohio and Michigan to allow correlation of stratigraphic units that have problematic chronological positions; and (4) to provide insights into the diagenetic processes inhibiting the study on their spatial and temporal relationships.
Figure 1.1 Upper Silurian and Devonian stratigraphic nomenclature of SW Ontario reflecting the positions of major facies changes and associated tectonostratigraphic controls due to intermittent movements of the forebulge region within present-day western Niagara Peninsula region and the remainder of SW Ontario (Brunton et al. 2017).
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1.3 Rationales and Approaches

This project employs a multidisciplinary approach integrating sedimentological, stratigraphic, paleontological and geophysical data to analyze the lithofacies distribution and to establish the regional sequence stratigraphy framework. It builds upon recent updates of Silurian studies in Ontario (Brunton et al. 2007; Brunton and Dodge 2008; Brunton 2009a, 2009b; Brunton and Piersol 2009; Brunton and Brintnell 2011; Brunton et al. 2012; Brunton et al. 2017) and New York State (Ver Straeten and Brett 2000; Brett et al. 2011). New data has been acquired through mapping and logging of all key cored wells, select cuttings and type sections to reconstruct regional cross-sections. The entire upper Silurian to Middle Devonian strata in SW Ontario (see Fig 1.1) has been studied in each of the cored holes and outcrops in order to understand their subsurface architecture and stratigraphic framework.

1.3.1 Carbonate and Evaporite

The targeted strata are composed predominantly of carbonates and evaporites. Carbonate rocks occur in two main forms: limestone (CaCO$_3$) and dolostone ([Mg, Ca]CO$_3$) (Flugel 2004). Carbonates are mainly intrabasinal in origin, primarily formed by biological activities, and susceptible to post-depositional chemical and textural modifications (see Tucker 2009). Unlike siliciclastic sedimentary rocks that are susceptible to mechanical dynamics, carbonate rocks are sensitive to environmental changes, such as variations in temperature, sunlight, water salinity, detritus influx that can influence biotic productivity, thus affecting carbonate sediment accumulation and precipitation. Textural relationships among skeletal or non-skeletal grains, mud content and cements can provide evidence for the depositional environments ranging from near-shore lagoons, to platform or ramp organic build-ups, shelf margin to slope and basinal settings (see Dalrymple and James 2010). Temperature, water depth, water energy, paleogeography, and basin geometry are the controlling factors on carbonate deposition, which can be reconstructed.
using detailed study of the carbonate microfacies distribution (Flügel 2004).

Evaporites are chemical sedimentary deposits that are mainly precipitated from sea-water in arid area where evaporation exceeds influx (Warren 2010). The principal evaporite minerals are anhydrite (CaSO$_4$), gypsum (CaSO$_4$ · H$_2$O) and halite (NaCl). Both subaerial and subaqueous evaporites have been found in sedimentary records worldwide and are discussed in detail by Kendall (1984), Schreiber et al. (1986) and Warren (1989, 2006, 2010). In the study area, syn-depositional evaporites are commonly interbedded with carbonates, indicating sea-level drawdown and water recharging cycles. Diagenetic evaporites are commonly replacive or displacive, distorting the original sedimentary structures (Warren 2010). Evaporite facies are grouped into three categories: mud-flat (sabkha) facies, shallow-water facies and deep-water facies (Dalrymple and James 2010). However, similar evaporite lithofacies can form in different depositional regimes. Therefore, identifying the origin of brines and location of barriers, together with the interpretation of associated carbonate facies, is the key to paleo-environment reconstruction. Evaporites are susceptible to dissolution after their formation and thus the paleo-topography would have been partly controlled by the dissolution and karsting of the underlying evaporites in SW Ontario.

1.4 Geologic Setting

Paleogeographically, the study area of SW Ontario straddles two Paleozoic sedimentary basins, the eastern margin of the Michigan Basin and the western margin of the Appalachian Foreland Basin (Fig 1.2). Physiographically, the study area represents the Western St. Lawrence Lowland region (Sanford 1968; Johnson et al. 1992).

The modern Michigan Basin is interpreted as a roughly circular, intracratonic basin centred in the lower peninsula of the Michigan State, USA (Allling and Briggs 1961; Howell 1990; Cercme
and Pollack 1991). It covers an area of 196,400 km$^2$ and attains its deepest deposit of approximately 5 km near the basin centre (Telford 1978). The Appalachian Foreland Basin contains $>11$ km thick sedimentary succession that records the intermittent orogenic phases spanning tens of millions of years and covers an area of 536,000 km$^2$ in eastern North America (Colton 1970; Ettensohn 2008). The preserved Paleozoic succession in Ontario is barely 1.4 km thick (Armstrong and Carter 2010). Situated between the foreland basin and the intermittently subsiding Michigan Basin, the Paleozoic strata blanket the Laurentian craton in a broad northeast-oriented zone referred to as the Algonquin Arch (Fig 1.2). Previous sequence stratigraphic studies on lower Silurian strata of SW Ontario have revealed that the Algonquin Arch can be best described as an ephemeral flexural forebulge that migrated from north to south through the Silurian (see Brett et al. 2004; Brunton 2009; Brintnell et al. 2009; Brunton et al. 2012). The Algonquin Arch is considered as a northeastern extension of the Findlay Arch in Ohio, and is separated by a local structural low called the Chatham Sag (Sanford 1985) (Fig 1.2 and 1.3). Both the orogenic activity at the eastern margin of North America and the locations of the basins and/or arches influenced the formation of sedimentary rocks within these two basins (Johnson et al. 1992).

The basement of the study area belongs to the Central Gneiss Belt of the middle Proterozoic Grenville orogeny (McLelland et al. 1996). It is located between two major crustal features, the Grenville Front Tectonic Zone and the Central Metasedimentary Belt Boundary Zone (Sanford et al. 1985; Easton and Carter 1995; Wallach et al. 1998; see Fig 1.2). These subsidiary structures within this crystalline basement formed under ductile to brittle-ductile conditions 1,100 Ma ago and dip moderately to gently to the E and SE (Sanford et al. 1985). The exact mechanisms of their influence on reactivation and passive fracture mechanisms are poorly constrained (ITASCA CANADA and AECOM 2011). The Paleozoic sedimentary sequences were deposited
nonconformably over the Precambrian basement, after an extended period of erosion representing up to 500 million years of exposure of the Grenvillian igneous and metamorphic rocks (Telford and Johnson 1984; Johnson et al. 1992). Sediment accumulations proceeded episodically from Cambrian to Carboniferous in response to several episodes of basin subsidence and arch uplift (Root and Onasch 1999). The Cambrian to Devonian strata preserved in SW Ontario dip at low angles (0.5°) towards the south in the Niagara region and towards the southwest in Bruce County, i.e. towards the depositional center of the two basins respectively. Two basement fracture systems (west to NW-trending set in the west and NE-trending set in the east) are consistently oriented across the region, corresponding to the Appalachian Orogenic phases (Fig 1.4) to the southeast (see Sanford et al. 1985; Ettensohn 1994; Root and Onasch 1999).

1.4.1 Michigan Basin

The current Michigan Basin has an elliptical outline, nestled against the southern margin of the Canadian Shield (Lavoie 2008) (Fig 1.3). The sedimentary rocks in the basin, largely Paleozoic and Mesozoic in age, reach a maximum thickness of 4,848 meters in the basin center located in the lower peninsula of the State of Michigan (Howell and van der Pluijm 1990). The Canadian Shield bounds the basin from northwest to the northeast, and the remainder of the basin is defined by the Wisconsin and Kankakee Arches to the southwest in Wisconsin, Illinois, Indiana and Ohio, and the Algonquin and the Findlay Arches to the southeast in western part of SW Ontario (Sanford et al. 1985). The strata of middle Cambrian to Upper Pennsylvanian are well represented throughout the subsurface in the lower peninsula of Michigan (Catacosinos and Daniels, Jr. 1991). Beneath the glacial sediments in Cenozoic, the subcrop of various formations forms a series of concentric patterns that outline the basin margin, i.e. the youngest near the center and the oldest at the margin (Ettensohn 2008).
Subsidence within the Michigan Basin occurred from middle Cambrian to Lower Jurassic periods (Sloss 1982, 1988; Howell and van der Pluijm 1990, 1999; Cercone and Pollack 1991). The continuity of sedimentation in the Michigan Basin was not consistent with that in the Appalachian Foreland Basin, due to the flexure and movement of the forebulge region in Ontario (Sloss 1988; Howell 1990; Ettensohn and Brett 2002). In addition, the subsidence history within the Michigan Basin itself was not laterally or temporally uniform either (Fisher et al. 1988). Several mechanisms have been proposed to explain the irregular subsidence history of the basin. Sanford (1985) argued that differential movement was caused by differential subsidence of individual fault-bounded megablocks (e.g., the Bruce Megablock to the northwest and Niagara Megablock to the southeast). Alternative hypothesized mechanisms include thermal contraction following development of an isolated “hot spot” (Sleep 1971; Nunn and Sleep 1984; Houseman and England 1986), metamorphic phase changes in the crust (Haxby et al. 1976; Middleton 1980; Ahern and Dikeou 1989; Hamdani et al. 1991), lithospheric stretching (McKenzie 1978; Klein and Hsui 1987), free thermal convection (Nunn 1994), and intraplate stress mechanisms (DeRito et al. 1983; Lambeck 1983; Howell and van der Pluijm 1990). Contention among these proposals centre mainly on the irregular basin subsidence rates (Sleep and Sloss 1978; Quinlan 1987; Howell and van der Pluijm 1990).

Though the proposed mechanisms are controversial, the pronounced factor is suspected as mantle plume activity that determined its saucer-like form through either localized lithospheric thinning (Mitrovica et al. 1989) or lithospheric drawdown by the cooling of underlying plume materials (Howell 1990; Kominz and Bond 1991). Alternatively, recent work shows that the Michigan Basin is the far-field of a forebulge region or might be a sub-basin of the Appalachian Foreland Basin that reflects the effects of craton-margin tectonics (see Quinlan 1987; Beaumont
et al. 1988; Root and Onasch 1999; Brunton and Brintnell 2011; Brintnell 2012; Brunton et al. 2012). In any case, the generally agreed basis of the basinal stratal distribution indicates an early subsidence starting from the Late Cambrian into the Early Jurassic (Sloss 1988), and during the late Silurian to Middle Devonian, the subsidence of the Michigan Basin reactivated with renewed tectonic loading associated with Salinic Orogeny (Boucot and Johnson 1967; Ettensohn and Brett 2002).

Since the early 1980s, various deep drilling projects and the resultant expansion of the stratigraphic database facilitated the division of the Michigan Basin stratigraphic succession into four genetic sedimentary sequences (defined by significant changes in the basin-subsidence patterns (Sloss 1988; Cercone and Pollack 1991 (Fig 1.4). The Proterozoic Midcontinent Rift System is considered as the basement of Michigan Basin (Smith 2002). From late Cambrian to Early Ordovician the Michigan Basin depositional units define an elongate trough, thickening to the east and west in the subsurface of southern Michigan and is completely removed in SW Ontario (Sleep et al. 1980. Johnson et al. 1993). During the Late Ordovician to Middle Devonian, the basin begun to separate from the previous pattern into a platform-like pattern (Coakley et al. 1994; Coakley and Gurnis 1995). Isopachs of the Upper Ordovician to lower Silurian sequence record no basin-centered subsidence, suggesting that the region possibly tilted eastward toward the rapidly subsiding Appalachian Foreland Basin (Quinlan and Beaumont 1984; Beaumont et al. 1988). Notably, the platform-like subsidence pattern may have lasted into Devonian (Allen and Armitages 2012). This phenomenon is attributed to regional-scale lithospheric deflection associated with subduction-related mantle flow, in response to the Taconic and Salinic orogenies (DeRito et al. 1986; Coakley and Gurnis 1995), in a manner similar to that of the Cretaceous Western Interior Basin (Beaumont et al. 1993). The episodic subsidence record of the Michigan
Figure 1.4 Phanerozoic Tectonic Cycles of Laurentia (North America) eastern margin during Paleozoic and western Eurasia margin during Mesozoic-Tertiary. Widths of the band represent relative intensity of tectonism during various tectonic events and farfield responses (after Sanford 1993). The study interval of strata was deposited in the locally Salinic Orogeny and initial phase of Acadian Orogeny.
Basin contrasts with the thermal-contraction subsidence model of the Illinois Basin to the south. The Illinois Basin appears to have a smoother Ordovician to Silurian subsidence history that is consistent with the reactivated loading in eastern Laurentia and the adjacent Appalachian (Ettensohn and Brett 2002). However, evidence of the significant subsidence reactivations corresponding to the subsidence episodes of the Michigan Basin is lacking (Sleep et al. 1980; Heidlauf et al. 1986; Watso and Klein 1989). During the Early and Middle Devonian, after the regional uplift and erosion in Lochkovian and possibly Pragian, the Michigan Basin started its inception of carbonates in late Early Devonian to Middle Devonian till the first phase of Acadian Orogeny of late Middle Devonian (Sanford 1993). In SW Ontario, orogen-derived siliciclastic sediments with minor carbonates were deposited in a seaway that covered the eastern margin of Michigan Basin during the following phase of Acadian Orogeny in Devonian (Johnson et al. 1992).

1.4.2 Appalachian Basin

As it is preserved today, the NE-trending Appalachian Foreland Basin is about 2,050 km long, covering nearly 536,000 km² from northern Alabama in the USA to southern Quebec in Canada (Colton 1970). Its elongate shape today reflects the structural influence of the Alleghenian Orogeny, which is the latest phase of the evolutionary history of the eastern margin of Laurentia in Permian (Ettensohn 2008). The distribution of Paleozoic strata filling the foredeep ranges from 600 m on its western flank to 13,700 m to its east (Sanford 1993). In general, the Paleozoic strata become thicker eastwards toward the Appalachian orogenic belt. The basin is bounded on the west by the Cincinnati, Findlay, and Algonquin arches (Fig 1.3), and on the east by metasedimentary, metavolcanic, and intrusive Precambrian and Paleozoic rocks of the Adirondack dome, Blue Ridge and New England Uplands (Colton 1970). Its current northwestern boundary in southeastern Ontario and southern Quebec is characterized by the up-dipping erosional limit of Paleozoic
sediments along the Laurentian and Frontenac arches of the Canadian Shield, whereas its southern boundary is transitional with the Black Warrior Basin.

Defined by tectonic setting and origin, the Appalachian Basin is a multistage or composite foreland basin (Dickinson 1974; Miall 1995) that largely formed in response to tectonic loading during four, nearly continuous orogenies on the eastern margin of Laurentia from the Early-Middle Ordovician transition through Permian time (Fig 1.4), including: 1) Taconic (Middle Ordovician – early Silurian); 2). Salinic (Silurian); 3) Acadian (Early Devonian – Mississippian); and 4). Alleghenian (Pennsylvanian – Permian) (Sanford 1993; Ettensohn and Brett 2002). The climax of these orogenic cycles coincides with major phases of basin subsidence and arch uplift, and influenced the sedimentary input into the region (Sanford et al. 1985).

The Caledonian orogenic cycles (Taconic and Salinic orogenies) in Ordovician–Silurian on the Appalachian margin of Laurentia represents closure of the Iapetus Ocean through subduction and collision involving island-arc complexes and peri-Gondwanan terranes during formation of Laurussia (Cook and Bally 1975; Ziegler 1989). In Eastern North America, the classic view holds that the Latest Silurian to early Middle Devonian was interpreted as an epierogenic interval during which no major orogenies occurred except for a regional uplift across the Silurian-Devonian (S-D) boundary (Sanford et al. 1985). Recent work shows that the Salinic Orogeny spans from early Silurian to Early Devonian, when movements of short-lived and ephemeral forebulges may have played an important role in the sedimentation pattern on the far-field of the Foreland Basin, and the Michigan Basin was interpreted as the Laurentia craton interior during Silurian-Devonian (see Brunton and Brintnell 2011; Brintnell 2012; Brunton et al. 2012).

The following orogenies, in contrast, reflect closure of the Rheic Ocean during collision with Gondwana and the related transpressional adjustment of peri-Gondwanan terranes to form the
super-continent Pangea. It occurred on the Appalachian margin of Laurussia during two orogenies - the earlier Acadian (including the Neo-Acadian), progressing from north to south, and the later Alleghenian, probably progressing from south to north in its later phases (Ettensohn 2008).

Each of these orogenic episodes and its associated lithospheric loading led to the asymmetric deepening of the trough-like foredeep proximal to the orogenic belt, forming the SE-trending set of joints and faults that are parallel to the Appalachian orogeny (Bell and Newman 2006).

1.5 Overview of the Latest Silurian–Middle Devonian World

The Silurian-Devonian world experienced periodic, exceptional sea-level highstands and global warm climates (Copper 2002). During the Silurian, the mega-continent Laurussia was formed by the welding of Laurentia-Greenland and Baltica-Avalonia along the Arctic-North Atlantic Caledonian suture (Ziegler 1988; see Fig 1.5), while the supercontinent Gondwana occupied most the southern hemisphere. A protracted eustatic sea level highstand and greenhouse climate, starting from early Silurian, promoted a peak time for global reef formation and distribution during Wenlock (Wood 1999; Copper 2002b). A drop in sea level during late Silurian (420-400 Ma) (Haq and Schutter 2008), interpreted as the closure of the Iapetus ocean between Laurentia and Baltica-Avalonia, resulted in the reduction of reefal diversity and uplifting of North America. Transgression of large shallow seas returned from early Emsian (late Early Devonian) into Frasnian (Late Devonian), which expanded the carbonate deposition and reefal fauna diversity (Copper 2002a). During Silurian-Devonian, the reef ecosystem was dominated by stromatoporoids and tabulate-rugose corals, with other contributors such as brachiopods, molluscs, trilobites and echinoderms (Wood 1999). The Devonian is also known as the “age of fishes”, marked by rapid diversification of both jawless and jawed fishes, with simple-jawed placoderms becoming the
dominant marine predators, especially in shallow marine and reef ecosystems (Long and Young 1995). Similar to marine realms, plants began colonizing terrestrial ecosystems in the Silurian, and their evolution and diversification accelerated during the Devonian. The development of spores led to the great diversity of trees and the first forest occurrence in Late Devonian (Algeo and Scheckler 1998).

1.6 Paleogeography of North America

During much of the mid-Paleozoic, eastern North America was in tropical latitudes (Fig 1.5), with a pronounced southerly drift that occurred during Late Silurian to Early Devonian (420 to 400 Ma) time after the collision of Laurentia and Baltica/Avalonia (Cocks and Torsvik 2011). Previous paleomagnetic and paleobotanic data indicate that the study area was positioned in a paleolatitude between 10°S and 15°S during early and middle Silurian and moved to the semi-arid tropical climate zone (20°S to 25°S) with high evaporation in the late Silurian with a clockwise rotation of 5–8° compared to today (Van der Voo 1982, 1988; Cocks and Torsvik 2011). During the Early Devonian, the study area was located at 30° to 35° south of the equator (Cocks and Torsvik 2011) and intermittently covered by inland seas. Therefore, the upper Silurian–Middle Devonian succession contains erosional and non-depositional gaps mainly due to marine regression, subaerial exposure and ephemeral tectonic uplift, resulting in an incomplete stratigraphic record (Johnson et al. 1992). Later, the tectonic plate moved northward and rotated counter-clockwise, returning to an equatorial or subequatorial position by the late Devonian (Cocks and Torsvik 2011). From the late Silurian to Middle Devonian, the eastern margin of the Michigan Basin was dominated by restricted carbonate successions with evaporite deposits, whilst the western margin of the Appalachian Basin received open marine deposits. In late Silurian,
Epeiric seas covering the two basins became possibly restricted by barriers of the clastic wedge to the south and barrier reef complexes around the Michigan Basin (Sanford 1969; Bartholomew et al. 2006). Previous studies indicate an uplift of the entire study area starting from the Silurian-Devonian boundary interval, initiating a retreat of the sea from the platform, and an erosional interval that lasted to Early Devonian (Sanford 1993; Johnson et al. 1992). It was followed by a return to narrow basin-centered subsidence continuing into Middle Devonian time (Sanford 1993).

In comparison to the Michigan Basin, the Appalachian Basin experienced a more open marine condition during the Late Silurian to Early Devonian. The loading of the continental margin, uplift and deformation of the Acadian Orogeny led to rejuvenation and reorganization of the foreland basin (Sloss 1988). By the early Middle Devonian, erosion resulted in progradation of clastic deposits into the basin from highlands with eventual infilling and gradationally more onto the craton, succeeding an open marine, high energy depositional environment pervasive during the in early Middle Devonian. The basin area was still located in the semi-arid, subtropical trade wind belt during Middle Devonian (Scotese 2003), although by the Late Devonian time relatively humid conditions returned (Caputo and Crowell 1985; Veevers and Powell 1987). From the Early to Middle Devonian, a delamination and slab loss (Van Staal and de Roo 1995; Tremblay and Pinet 2005) during the Helderberg transition of dextral, oblique convergence between Avalonian terranes and Laurussia (Trupe et al. 2003) has been interpreted to be responsible for sediment accumulation (Pragian, Emsian, and Eifelian stages) at western margin of the Appalachian Basin, forming a vertical succession of marine siliciclastics and detrital carbonates originated from the migrating clastic wedge (Sloss 1963).
Figure 1.5 Paleogeography map of the North America during late Silurian to Early Devonian. Red box shows the location of study area. (A): late Silurian (420 Ma). North America located between equator and 25° south, the major craton of which is covered by epeiric seas. The study area received deposits within both the Appalachian Foreland Basin and the Michigan Structural Basin. (B): Early Devonian (400 Ma). North America located near the 35° south latitude and was regionally uplifted. (Provided by Deep Time Map™ and modified from Cocks and Torsvik 2011).

1.7 Materials and Methods

In this study, cores, outcrops, thin sections and geophysical data were used for lithofacies analysis, stratigraphic correlation, and interpretation of depositional environments and diagenetic history. Core data obtained via direct observation include lithologic and faunal descriptions and characterization of major contacts of the stratigraphic units. Thin sections made from cores are used for describing the important petrological characteristics and key stratigraphic contacts. Lateral lithofacies changes in lithology and stratigraphic architecture are examined in detail in available outcrops, in which the visibility of formational contacts tends to be accentuated by
weathering and sequence boundaries laterally. Geophysics data and cuttings are incorporated to fill between the gaps in none-cored wells. By integrating these methods, the stratigraphic units can be reliably defined and correlated in the study area.

1.7.1 Core Analysis

The bulk of this study is based on lithological data obtained from cores with gamma-neutron-density logs of 38 cored wells (Table 1.1) and selected regional oil/gas wells drilled in southern Ontario. Cores from southern Ontario are stored at the Oil, Gas, and Salt Resource Library in London, Ontario and at the Ontario Geological Survey in Sudbury, Ontario. Well identification data and distribution are shown in Table 1.1 and Figure 1.6. Wells were first selected and observed from the easterly Bruce County and Huron County in the Michigan Basin, then eastward into the Onondaga Escarpment in southern Ontario, and southward into Lake Erie across the interpreted locations of Algonquin Arch and Chatham Sag. Cores extending from the top Dundee Formation to the base of the Bass Island Formation were targeted for analysis. Cores that cross the Silurian-Devonian boundary were selected in priority to log and to characterize the features of this regional unconformable contact.

Cores have been logged in detail using the following general characteristics: 1) carbonate rock types; 2) presence and proportion of fossils or non-skeletal constituents; 3) porosity types; 4) depositional structures; and 5) diagenetic fabrics, etc. Major mappable lithofacies have been differentiated and described in each area accompanied by sketches and photographic illustrations of key cores during logging for further interpretation of deposition regimes. Major formational contacts and/or deposition hiatuses have been described and characterized as regional stratigraphic correlation marker horizons.
Figure 1.6 Well locations. Geologic map of southwestern and parts of south-central Ontario showing the locations of oil and gas wells and groundwater exploration or karst wells that penetrate the stratigraphic interval of interest. This region is part of the western St. Lawrence Lowlands, comprising Paleozoic-dominated bedrock and overlying Quaternary sediments. The red line outlines the study area for the upper Silurian–Middle Devonian sequence stratigraphy and bedrock groundwater aquifer mapping project. Numbered red dots represent existing cored holes through all or parts of the stratigraphic interval of interest (see Table 1.1). Orange-yellow dots represent oil and gas wells with geophysical records from Middle Devonian Dundee Formation down to Upper Silurian Salina Group strata. The Paleozoic bedrock geology and Grenville Province geology units and colours are from Armstrong and Carter (2010). The bathymetry for the Great Lakes was created using the Great Lakes Bathymetry data, courtesy of the National Oceanic and Atmospheric Administration (2007). (After Sun et al. 2014)
Table 1.1 List of cored wells. List of key drill holes that are cored through all or parts of the Upper Silurian through Middle Devonian strata under investigation across SW Ontario.

<table>
<thead>
<tr>
<th>Drill Hole</th>
<th>Licence - Location</th>
<th>Well Name</th>
<th>Easting (m)</th>
<th>Northing (m)</th>
<th>Remarks or Reference</th>
</tr>
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<td>4713938.051</td>
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1Number refers to location shown on Figure 1.5.
2Abbreviations: BH, borehole; DDH, diamond-drill hole; MTO, Ministry of Transportation Ontario; OGS, Ontario Geological Survey; OPG-DGR, Ontario Power Generation Deep Geologic Repository; OGSR, Oil, Gas and Salt Resources Library; SVCA, Saugeen Valley Conservation Authority.
3Universal Transverse Mercator (UTM) co-ordinates provided using North American Datum 1983 (NAD83) in Zone 17.
A total of 117 hand samples were collected, representing different lithofacies from key cores in Niagara Peninsula and Bruce and Huron Counties. Petrographic analyses of 98 thin sections, including the examination of grain and fossil types, porosity, depositional texture and diagenetic fabric enables microscopic properties of carbonate facies to be identified. Photomicrography have been undertaken using a Leica Microscope Z16 APO in the paleontology lab, University of Western Ontario.

1.7.2 Outcrop Sections

Several outcrop sections in southern Ontario were measured and logged, ranging from upper Silurian to Middle Devonian, located in quarries in Niagara area, along the Onondaga Escarpment, from London to Ingersoll, as well as road-cuts and quarry sections from Sarnia to (Table 1.2). Most old exposures documented in Derry Michener Booth et al. (1989) have been weathered severely. Some of the fresh exposures are found in recent quarry operations and have been integrated in the regional stratigraphic correlations. Key outcrop sections examined in this study include the followings:

1). The type section of the Bertie Formation in SW Ontario, Ridgemount Quarry, west of Fort Erie, Bertie Township (see Winder 1961; Haynes and Parkins 1992; Armstrong and Dodge 2007). The Ridgemount Quarry North exposed 14.25 m of section of the Bois Blanc and Bertie formations. All five members of the Bertie Formation and the contact with the overlying Bois Blanc Formation are exposed. The Ridgemount Quarry South is an abandoned quarry, partly covered by water. It exposes 10.7 m of section ranging from the upper Bertie to Middle Devonian Onondaga formations at its north side. Glauconite occurs commonly at the Silurian-Devonian boundary in the quarry.
2). The Hagarsville Quarry is an active quarry located at the northeastern outskirts of Hagarsville, east of Plank Road. This quarry exposes 16 m of the upper Silurian Bertie Formation and the Lower Devonian Bois Blanc Formation including the sandstones of Springvale Member.

3). The Nelson Quarry is an intermittently operational quarry located 1.5 km south of Clanbrassil, town of Haldimand. The upper Silurian Bertie/Bass Islands formations and the Lower Devonian Bois Blanc are exposed, including 2.2 m of the Oriskany Sandstone and 4.4 m of the cherty limestone of the Bois Blanc limestone. The sandstone of Oriskany Formation is also exposed in the Cayuga Quarry, but access cannot be obtained from the operator. Therefore, the Nelson Quarry is the only available outcrop exposure of the Oriskany sandstone at the S-D contact.

4.) The Dunnville Quarry is located 1 km south of the town of Byng, currently inactive, and exposes the Bertie, Springvale Member, and the Bois Blanc Formation. The S-D unconformity is sharp, marked by folding and faulting in the underlying Bass Islands Formation, visible in the south wall of the quarry. The sandstone of Springvale Member (Bois Blanc Formation) is gluconitic and phosphatic, which is separated from the overlying fossiliferous Bois Blanc limestone by a smooth surface.

5). The Formosa Road-cut section is on Bruce County Road 12, about 3 km north of Formosa. The Formosa Reef facies (upper Amherstburg Formation) is exposed in the 48-m long road-cut section, consisting of abundant tabulate corals (*Favosites*) and tabular stromatoporoids. Rugose corals, brachiopods, bryozoans, trilobites, and crinoid columnals/ossicles are also locally abundant within the reefal framestones. Stromatoporoid forms are very well preserved with the original aragonitic skeleton inverted to pseudomorphic calcite (Tsujita et al. 2001).

6). The McGregor Quarry is a large active quarry in the Anderdon Township, 10 km northeast of Amherstburg. This quarry exposes 32 m of the Lucas Formation (Detroit River Group).
Evaporitic dolostone facies are present in the lower 19 m, and the upper 13.65 m represents the Anderdon Member that consists of pure fossiliferous limestone. The overlying Dundee Formation is partly exposed in the quarry, but too high to examine.

7). The Dundee Formation and its disconformable contact with the underlying Lucas Formation is well exposed in the St. Mary’s Quarry, St Mary’s (Armstrong and Carter 2006, 2010).

**Table 1.2** List of outcrop locations and formations represented in targeted formations outcrops used in this study. Fm=Formation

<table>
<thead>
<tr>
<th>Outcrop Section</th>
<th>UTMe</th>
<th>UTMn</th>
<th>Formation(s) Represented</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ridgemount Quarry North</td>
<td>662290</td>
<td>4753935</td>
<td>Bois Blanc Fm and Bertie Fm</td>
</tr>
<tr>
<td>Ridgemount Quarry South</td>
<td>662676</td>
<td>4753308</td>
<td>Onondaga Fm, Bois Blanc Fm and Bertie Fm</td>
</tr>
<tr>
<td>Thunder Bay east</td>
<td>660179</td>
<td>4748511</td>
<td>Clarence and Edgecliff Mbr (Onondaga Fm)</td>
</tr>
<tr>
<td>Morgan’s Point</td>
<td>634873</td>
<td>4746404</td>
<td>Edgecliff Mbr (Onondaga Fm)</td>
</tr>
<tr>
<td>Dunnville Quarry</td>
<td>612890</td>
<td>4748675</td>
<td>Bois Blanc Fm; Bertie/Bass Islands Fm</td>
</tr>
<tr>
<td>Cayuga Quarry</td>
<td>586845</td>
<td>4755430</td>
<td>Bois Blanc Fm; Oriskany Fm; Bertie/Bass Islands Fm</td>
</tr>
<tr>
<td>Hagarsville Quarry</td>
<td>579354</td>
<td>4756950</td>
<td>Bois Blanc Fm; Springvale Mbr (Bois Blanc Fm); Bertie/Bass Islands Fm</td>
</tr>
<tr>
<td>Selkirk-Hemlock Creek</td>
<td>587372</td>
<td>4741786</td>
<td>Moorehouse Mbr (Onondaga Fm)</td>
</tr>
<tr>
<td>Shore west of Hemlock Creek</td>
<td>586930</td>
<td>4740462</td>
<td>Dundee Fm and Onondaga Fm</td>
</tr>
<tr>
<td>Norfolk Quarry</td>
<td>566152</td>
<td>4740270</td>
<td>Dundee Fm and Onondaga Fm</td>
</tr>
<tr>
<td>Nelson Quarry</td>
<td>623136</td>
<td>4787636</td>
<td>Bois Blanc Fm; Oriskany Fm and Bertie Fm</td>
</tr>
<tr>
<td>Amherstburg Road Cut</td>
<td>482210</td>
<td>4882825</td>
<td>Formosa Reef (Amherstburg Fm)</td>
</tr>
<tr>
<td>MacGregor Quarry</td>
<td>333660</td>
<td>4686693</td>
<td>Lucas Fm</td>
</tr>
</tbody>
</table>
1.7.3 Regional Cross-Sections

Stratigraphic cross sections, based on mean sea level as datum, have been constructed using the formational tops and regionally correlative marker beds as correlation horizons based on lithological characteristics and macrofossil content seen in cores, integrated with geophysics data and cuttings. Well-line/geophysics logs are measurements of the physical characteristics of the rock formations. For the upper Silurian-Middle Devonian successions in SW Ontario, gamma-ray and neutron logs are the most useful and best recovered data for regional stratigraphic correlation. A gamma-ray log is a measurement of the natural gamma ray emissions radiated by potassium-rich feldspars that are the common components in shales or feldspar-rich sandstone. In carbonate successions, a high response in gamma-ray log usually indicates the presence of shale layers. Neutron logs are used primarily for measurement of relative porosity of subsurface rocks. It is a method to delineate ratio of the volume of pore spaces to solid rocks filled by hydrocarbons or formational waters. Gamma-ray and neutron logs are usually used together in the subsurface mapping programs to provide information about shale layers and any porous horizons in the wells, especially where no core is recovered. Previous studies have, however, shown difficulties in using the gamma ray log to distinguish formational contacts in the targeted carbonate packages lack shale marker beds. Hence other methods have been employed, such as packer testing and heat pulse flow metering; downhole cameras; optical and acoustic televviewer logging; instrumentation of holes with ports to collect waters at flow zones and collection of cuttings.

1.8 Organization of the Thesis

Based on the objectives and methods, the rest of the thesis are formed up by the following nine parts. Chapter 2-4 describes the lithofacies, stratigraphy and regional correlation of the Silurian Bass Islands Formation and Bertie Formation, including the underlying Salina G Unit. In
chapter 4, the paleokarstic features and paleokarst systems developed within the Bass Islands Formation below the S-D contact in cored wells are delineated. Chapter 5 documents the S-D unconformities in SW Ontario and its significance for the study of bedrock aquifers and the corresponding tectonic evolution of the SW Ontario. Chapter 6-8 provided descriptions of the lithofacies and stratigraphy of the Devonian Onondaga Formation and Dundee Formation in Ontario. In Chapter 8, a sequence stratigraphic approach has been employed to establish a regional stratigraphic framework for better correlation of the Onondaga Formation into the other Devonian units in SW Ontario, Michigan and Ohio. Chapter 9 provides a summary of the key findings of this thesis and hold discussions on a depositional history of the upper Silurian-Middle Devonian carbonate rocks in SW Ontario. Legends for all lithologic figures are listed in Figure 1.7.
Figure 1.7 Legends for lithology figures in Chapter 2 – Chapter 8.
CHAPTER 2. Lithofacies Descriptions and Interpretations of the Silurian Bass Islands Formation and the Bertie Formation, Southwestern Ontario

2.1 Introduction

The youngest Silurian deposits in SW Ontario belong to the Cayugan Series (Pridoli), which is divisible into two major sequences: the lower evaporite, dolomite and shale succession of the Salina Group, and the upper Bass Islands Formation on the Laurentian craton (Michigan structural Basin) and the Bertie Formation on the western margin of the Appalachian Foreland Basin in Niagara area and New York (Sanford 1969; Sparling 1970; Haynes 1992). The Bass Islands and Bertie formations are thought to be equivalent by both immediately underlying the Silurian-Devonian contact and both displaying lithologically similar peritidal dolostones (Fig 2.1) in the Michigan and Appalachian basins respectively (Sanford 1969; Johnson et al. 1992). Their stratigraphic nomenclatures in SW Ontario are largely based on lithostratigraphy (see Beards 1967; Liberty and Bolton 1971; Telford and Tarrant 1975a, 1975b; Armstrong 2017). The biostratigraphy is not well constrained due to poor conodont and other faunal recovery (Sanford 1968; Uyeno et al. 1982).

The Bass Islands Formation hosts the deepest bedrock portable water aquifers in Ontario, but the previous studies on the formation and distribution of these aquifers are limited (Carter and Brunton 2011; Carter 2012; Brunton and Dodge 2008). Therefore, a stratigraphic reexamination and a detailed study of the paleokarstic systems within the Bass Islands/Bertie formations are necessary to refine the understanding of the regional architecture of the upper Silurian carbonate/evaporite successions in Ontario, in order to delineate geological controls on the formation and distribution of the bedrock aquifers across from Lake Erie into Lake Huron. The
overlying Silurian-Devonian unconformity in SW Ontario represents a major depositional hiatus – the late Silurian carbonate rocks display significant erosional features in the form of paleo-karst (e.g., solution-widened joints, sinkholes and collapse breccias, in places filled with variable siliciclastic materials) below the unconformities in both subsurface bedrock and scattered outcrops (see Kobluk et al. 1977; Brunton and Dodge 2008). The overprint of diagenetic features, such as replacive anhydrite nodules and severe dolomitization, also inhibits the interpretation of the depositional history of these youngest Silurian deposits in SW Ontario. In Oxford and Huron counties, the thick glacial deposits have hindered the observation of the Bass Islands Formation in outcrops, too. Therefore, 38 cored holes have been selected for this study to delineate the lithological and sedimentological features of the Bass Islands Formation in the subsurface of SW Ontario.

2.2 Nomenclature

In recent years, with increased subsurface observations incorporated with core logging, seismic survey imaging and geophysical logs, a revised subsurface stratigraphic terminology chart for SW Ontario (Fig 1.1) has been proposed by Brunton et al. (2017). Sequence stratigraphic/allostratigraphic methods have also been employed to establish interpretative genetic models across parts of eastern North America (Johnson et al. 1992; Sanford 1993; Brett 1998; Brett et al. 2011). Challenges remain, however, to associate merging outcrop-based surface nomenclatures with subsurface terminologies. The subsurface terminology used for this study is shown in Fig 2.2 and 2.3, which follows Armstrong and Carter (2010) and Rickard (1975, 1984, 1985), with minor modifications.
Figure 2.1 Generalized lithologies of Bass Islands Formation and Bertie Formation, SW Ontario. See Fig 1.7 for legends.
**Figure 2.2** Nomenclatures of the Salina Group and the Bass Islands Formation in eastern Michigan, northern Ohio and SW Ontario.
**Figure 2.3** Nomenclatures of the Salina Group, the Bertie Formation and the Bass Islands Formation in western New York and SW Ontario.
2.2.1 Bass Islands Formation

In the Michigan Basin, the Bass Islands Formation is the topmost Silurian strata in both State of Michigan and SW Ontario (Alling and Briggs 1961; Sanford 1969; Sparling 1970; Johnson et al. 1992; Tomastik 1997). The lithology is characterized as brown to light grey, variably laminated and thrombolitic, mottled, argillaceous or bituminous dolostone with very common breccias, evaporite interbeds and extensive paleokarstic features throughout the formation (Alling and Briggs 1961; Brunton and Dodge 2008; Armstrong and Carter 2010; Sun et al. 2016). It occurs beneath younger formations along western Lake Erie and subcrops along the eastern margin of Michigan Basin in Bruce, Wellington, Perth and Brant counties (Sanford 1969; Derry Michener Booth et al. 1989; Johnson et al. 1992). Its eroded cuesta edge is poorly exposed, with only one outcrop section has been described along the Saugeen River by Liberty and Bolton (1971). Haynes (1991) has reported the presence of Bass Islands Formation in Cayuga, Dunville and Hagarsville quarries, which have been identified as upper Bertie Formation in previous studies (Caley 1946; Liberty 1966; Liberty and Bolton 1971; Telford and Tarrant 1975a, 1975b).

Studies of the Bass Islands started in Ohio and Michigan in the early 1900s (Lane et al. 1909). The type section of the Bass Islands Formation is designated on Bass Islands, Ottawa County, north-central Ohio, USA (Lane et al. 1909; Winder 1961; Alling and Briggs 1961; Sparling 1970). A modern and significant study was made by Sparling (1970), who divided the type Bass Islands Formation in Ohio into the Greenfield, Tymochtee, Put-in-Bay and Raisin River members in ascending order. The lower two members are roughly equivalent to the Salina Group in Michigan and Ontario, and the upper Put-in-Bay and Raisin River dolostone members in Ohio are thought to be correlated to the Bass Islands Formation in Ontario (Alling and Briggs 1961; Sparling 1970; Haynes 1991; Johnson et al. 1992). Sparling (1970) reported that the gastropod *Goniophora dubia* (Hall) is abundant in the Put-in-Bay Member, and he tentatively placed it within the *Goniophora*
prosseri zone (Pridoli). The overlying Raisin River Member were designated to the *Whitfieldella prosseri* zone, and the locally diagnostic *Whitfieldella prosseri* zone overlaps with the *Goniophora prosseri* zone for about 15.5 m in the Put-in-Bay and Raisin River members in Detroit salt shaft, Ohio (Alling and Briggs 1961; Sparling 1970).

The Bass Islands Formation has a complicated nomenclatural history in Ontario. Williams (1919) was the first to recognize a Cayugan unit in a small quarry at Innerkip (east of Woodstock), and he considered it the dolostone in the Raisin River Member correlative to the Waterlime Group in New York. Caley (1946) recognized the Bass Island Formation in the Windsor-Sarnia area, although he retained the unit name Bertie-Akron used for the units in the Appalachian Basin. In the 1930s, Michigan Reports adopted the formal name “Bass Islands Formation” (Liberty and Bolton 1971) based on the work of Sanford and Brady (1955) who changed the name Bertie-Akron to Bass Island in the supplement materials to Caley’s reports. In subsequent works, this unit was referred to as the “Bass Island” in Ontario, and formalized as the “Bass Islands” in Ontario by the Geological Survey of Canada (Poole et al. 1967). All the publications in Ontario maintained this formation name in recent years for the uppermost Silurian dolostone unit (e.g. Telford and Tarrant 1975; Haynes and Parkins 1991; Haynes 1992; Johnson et al. 1992; Sanford 1993; Armstrong and Carter 2010).

In their type region, the Put-in-Bay and Raisin River members are recognized as partly brecciated, variously laminated, argillaceous dolostone and massive, finely-crystalline, dense dolostone (Alling and Briggs 1961; Sparling 1970; Tomastik 1997; Harrison et al. 2009). The contact of Put-in-Bay and Raisin River is laterally discontinuous and one observable difference is that the Put-in-Bay dolostone is more brecciated than that of the Raisin River Member. Thus Sparling (1970) suggested that the Put-in-Bay Member is a locally brecciated lithofacies of the
Raisin River Member and used the term “Bass Islands” to represent these two similar members in Ohio. In Ontario, these two members are not readily distinguishable in either outcrops and subsurface. In this thesis, therefore, the name “Bass Islands Formation” of Liberty and Bolton (1971) is used to refer to all the topmost Silurian dolomite strata overlying the evaporitic dolostone of Salina G Unit in SW Ontario.

The thickness of the Bass Islands Formation ranges from 10.5 m to 91.5 m on a regional scale, but may reach 150 m in parts of the subsurface in Essex and Bruce counties (Sanford 1969). The Bass Islands Formation consists of various lithofacies that change rapidly and drastically both vertically and laterally based on subsurface core loggings. Its characteristic laminated and argillaceous dolostone with common evaporite minerals indicate a peritidal to sabkha depositional environment, with various levels of paleokarstic breccias. Fossils are very rare. Ostracods, brachiopods, bivalves and gastropods can be sparsely present locally near the top Bass Islands Formation (Haynes 1992) as seen in cores from Huron and Essex counties. Sedimentary structures including horizontal or wavy parallel laminations, desiccation cracks, tepees, cross beddings, rip-up clasts, ball-and-pillow deposits and vertical brecciation pipes are very common.

In Bruce County where the Bass Islands Formation is exposed, the lower contact with the Salina Group is arbitrarily defined by the presence of peritidal-restricted lagoon deposits of microbialites or buff dolomites with common bituminous streaks overlying the reddish or greenish grey shale or shaly dolostone (Liberty and Bolton 1971). However, this relationship seems to represent the contact of the Salina F Unit and the Salina G Unit instead of the contact of Salina Group and the Bass Islands Formation. In the subsurface, the uppermost Salina Formation is placed at the highest appearance of gypsum/anhydrite nodules below the laminated shale or shaly dolostone (Haynes 1992; Armstrong and Carter 2010; Sun et al 2014; 2016). This is sharp and has
a high response in gamma-ray log. At the base of the Bass Islands Formation, the attribution of the dolomitic shale or shaly dolostone unit thins from 3 m in the Appalachian Basin westward to a few centimetres into the Laurentia craton interior. The upper contact of the Bass Islands Formation is featured by a regional erosional surface that marks the S-D contact. Sandstones of the Devonian Oriskany Formation or Springvale Member of the Bois Blanc Formation locally overlie the S-D contact in the Niagara area. Where these siliciclastic sandstone units are absent, the cherty carbonates of the Bois Blanc Formation overly the S-D contact uncomfortably.

2.2.2 Bertie Formation

As an Appalachian Foreland Basin unit, the Bertie Formation is considered equivalent to the Bass Islands Formation of the Michigan Basin and the Bertie Group of New York (Armstrong and Carter 2010). The type section of the Bertie Formation is in Bertie township (Winder 1961), which consists of Oatka, Fiddler Green, Scajaquada and Williamsville members in ascending order. In Ontario, most of the Bertie Formation is exposed along the Onondaga Escarpment in Niagara Peninsula area (Derry Michener Booth et al. 1989; Haynes and Parkins 1992; Brett et al. 2004; Sun et al. 2016), including quarries near Fort Erie, Port Colborne, Cayuga, Hagersville, Norfolk and Nelson. Nomenclature of the Bertie Formation has undergone significant changes since its early recognition in Bertie and Cayuga Townships, Niagara Peninsula by Chapman (1864) who correlated the “Bertie or Cayuga dolomite” to the poorly defined “Waterlime Group” in New York. The name “Bertie Waterlime” was used by New York geologists in the early 1900s for the whole upper Silurian unit overlying the Salina Group, until Grabau and Scherzer (1909) identified a separate unit overlying the Bertie Waterlime in Akron, New York, which they named as “Akron dolostone”. Chadwick (1917) recognized the Akron dolostone near Buffalo and subdivided the underlying Bertie Waterlime into four members, including, in ascending order, Oatka, Falkirk,
Scajaquada and Williamsville. Problems remain in New York on 1) whether the Akron Formation in western New York is equivalent to the Cobleskill Formation in eastern New York (Brett et al. 2000); 2) whether Bertie is of formational or group status; and 3) whether the Akron is a member within the Bertie Group/Formation. In New York, recent workers treat the Bertie as a group (subdivided into four formations: the Oatka, Fiddler Green, Scajaquada and Williamsville), which is overlain by the Akron-Cobleskill Formation (e.g. Brett et al. 2000; Ciurca Jr. 2011).

Williams (1919) was the first to map and establish the Bertie Formation in Ontario. He recognized the Akron dolostone in the Dunnville area and identified a unit overlying the Akron Member as the Raisin River dolomite of Grabau’s (1909) Bass Islands Formation in the Innerkip quarry. This contact relationship was confusing and contradicting with the presumably studies showing the equivalence of the Bass Islands and Bertie formations. Outcrops become sparse westward. The following major work was conducted by Caley (1941, 1943, 1946), who used the term “Bertie-Akron Formation” for the upper Silurian strata and retained the same name in the Sarnia-Windsor area, although he recognized the Bass Islands units in adjacent Michigan and Ohio. Caley did not recognize the Akron unit in the Niagara Peninsula, and this resulted in downgrading the nomenclature of Akron from a formational status to a member of the Bertie Formation. Hewitt (1960) mapped these uppermost Silurian units eastward along the Onondaga Escarpment, in which he noticed a distinct “mottled dolostone” of the Akron member underlying the Devonian units from Fort Erie to Port Colborne. The mottled dolostone of Akron pinches out westward and is replaced by the westward-thickening unit of the light buff, laminated dolostone, regarded as the Bass Islands Formation (Eckert 1976; Haynes and Parkins 1992). The term “Bertie Formation” has been widely used by Ontario geologists since then (e.g. Telford and Tarrant 1975; Johnson 1986; Johnson et al. 1992; Armstrong and Carter 2006, 2010; among other government reports).
Since Williams (1919), the location and nature of the contact of Bass Islands and Bertie have not been addressed. The contact location has been tentatively set at the border of the Simcoe by Telford and Hamblin (1980) and the town Dunnville (Telford and Tarrant 1975), both of which seem arbitrary because no outcrops were exposed. The petroleum industry in Ontario treats the Bass Islands and Bertie formations as two roughly equivalent units (Derry Michener Booth et al. 1989; Armstrong and Carter 2010), whereas Haynes and Parkins (1992) observed that the Bass Islands facies overlies the Bertie Formation in Dunnville to Hagarsville area.

The Bertie Formation in Ontario, as defined by Haynes and Parkins (1992) and used in this study, comprises five members: the Oatka, Falkirk, Scajaquada, Williamsville, and Akron, in ascending order. In Ontario and adjacent western New York, the Bertie Formation consists of a 16–18 m thick cyclic successions of buff grey, argillaceous or bituminous, thinly laminated or massive bedded dolostones, shaly dolostones and shales. All the five members can be distinguished in outcrops of the Niagara Peninsula area. There have been many stratigraphic overviews for the Bertie Formation in Ontario and New York (e.g. Rickard 1969, 1975; Craft 1964; Treesh 1972; Ciurca 1990, 2005, 2011; Belak 1980; Tollerton and Muskatt 1984; Haynes 1991; Hayens and Pakins 1992), based on which a summary of the five members is given below (see Fig 2.1).

**Oatka Member.** The lowest unit in the Bertie Formation, which comprises upto 5-m thick, greenish grey, wavy or irregular bedded, interbeds of shale and shaly dolostone. Fossils are very rare and 1-2 cm thick, coarsely crystalline evaporite interbeds have been reported by Hayens (1991). The lower contact is not exposed in outcrops (e.g. Ridgemount Quarry) and is considered as being marked by the highest occurrence of evaporite nodules is the Salina Group. The shale of Oatka Member forms a relatively impermeable layer that controlled the horizontal bedrock groundwater
flow zones within the overlying Bertie Formation (Hayens and Parkins 1992; Brunton and Dodge 2008). Stratigraphic position of this shale unit is uncertain — it represents either the top Salina Group or the basal Bertie Formation (see discussions in Ciurca 1990; Haynes and Parkins 1991).

_Falkirk Member._ The Falkirk Member is 5–6 m thick in the Fort Erie-Welland area (Armstrong 2017), approximately correlative to the Fiddlers Green Formation of the Bertie Group in New York (Craft 1964; Brett et al. 2000). It is characterized by medium to dark brown, hard, dolomitic cryptalgal boundstone or thrombolitic mounds with abundant argillaceous stylo-seams. In its reference outcrop region from Hagarsville Quarry to Ridgemount Quarry, the Falkirk Member overlies the greenish Oatka shale or clayey dolostone by an undulatory surface (Armstrong 2017). Individual subunit is not regionally continuous (Hayens and Parkins 1992). Breccias are very common in drilled cores and are considered as either tsunamiite (Ciurca 1990) or collapse structure caused by dissolution of salts of Salina Group (Hayens 1991). Brachiopods (*Whitfiedella*), ostracods and eurypterids (*Eurypterus remipes remipes*) have been reported in the Falkirk Member in western New York State (Ciurca and Hamell 1994). Mehrtens and Barnett (1977) have determined the conodont ranges from late Silurian to Early Devonian by the finding of *Ozarkodina remscheidensis remscheidensis* (Ziegler). Uyeno et al. (1982) also found this long-range species within the top 20 cm of the Falkirk Member.

_Scajaquada Member._ The succeeding Scajaquada Member is 1.4–2.3 m thick, dark grey to greenish grey, argillaceous or shaly dolostones or thin-bedded shales. Selenite discs or salt hoppers are locally common (Hayens and Parkins 1992), and fossils are lacking (Eckert 1976). It overlies the Falkirk Member by a sharp and undulatory surface. This shale unit is present continuously in
the Niagara Peninsula and in adjacent Buffalo area, but cannot be readily recognized west of Hagarsville area (Hayens 1991).

**Williamsville Member.** The Williamsville Member consists of pale-grey to light grey, fine-grained argillaceous dolostones interbedded with clayey dolomudstones. It is 2.3 m thick in Ridgemout Quarry, but thins to 1 m near Cayuga area in Ontario (Ciurca 1990; Hayens and Parkins 1992). Conchoidal fractures are present (Ciurca 2005, 2011). This member is known for its well-preserved specimens of eurypterids (*Eurypterus remipes lacustris*) (Ciurca 1974; Eckert 1976) and brachiopods (*Eccentricosta*) in western New York (Ciurca 1990; Brett et al. 2000). Recently, a fish fossil (*Nerepisacanthus denisoni*) was discovered as the earliest occurrence of this species by Burrow and Rudkin (2014). Its basal contact with the Scajaquada Member is regionally gradational and shows a decline in response to gamma ray log (Armstrong 2017).

**Akron Member.** The Akron Member consists of grey, massive to wavy bedded, vuggy dolomudstone with buff-grey common mottles. It is 5 m thick at Buffalo and thins to 2.5 m at Cayuga and to 1 m at Hagarsville, pinching out westward (Rickard 1975; Hayens and Parkins 1992; Brett et al. 2000). The basal Akron Member is weakly mottled with common wavy layers marked by coalesced carbonaceous seams. The occurrence of the characteristic buff motting of the grey Akron Member marks the basal contact with the underlying Williamsville, and the intensity of buff motteling increases in the upper Akron with common fenestrae vugs. Preservation of Ostracods (*Leperditia*) and molds of brachiopods and corals occur at the top Akron Member in Ridgemount quarry (Eckert 1976). It underlies the Devonian sandstones in Ridgemount to Port Colborne area, while in Cayuga to Dunville quarries it underlies a 5-m thick dolostone unit that is considered to
be the Bass Islands Formation by Eckert (1976) and Hayens and Parkins (1992). Desiccation features including tepees, degassing structures, polygonal cracks and mud chips are common within this unit. Small cavities infilled by Devonian sands or cherty dolostones are present locally in the up-section below the S-D contact. Belak (1985) showed that the dolomitic limestone of the Cobleskill Formation in eastern New York State is equivalent to the Akron Member although it is less diagenetically altered.

2.3 Area Divisions

In order to map the spatial distribution and regional lithofacies variations in the Bass Islands and Bertie formations, cored wells were selected based on the presence of this geologic unit for further analyses (see Table 1.1). Detailed core logging and description have been performed in three depositional areas (Fig 2.4), including northeastern margin of the Michigan Basin (area 1), southeastern margin of the Michigan Basin transitional into the northern Ohio (area 2) and northwestern margin of the Appalachian Foreland Basin (area 3). These three depositional motifs are confined roughly by the interpreted paleo-topographic highlands along the northeastern-southwestern trending axis of the forebulge region (Sanford 1969; Johnson et al. 1992; Ver Straeten and Brett 2000; Ettensohn and Brett 2008). Sanford (1969) suggested variable paleogeographic settings for the upper Silurian units in the Michigan Basin and Appalachian Basin of SW Ontario, which overlaps with portions of our study area. In previous studies, the Bass Islands Formation in Essex County, however, was not differentiated from the other two areas.

Detailed observation of lithology, fossil composition and sedimentary structures are examined, supported by microlithofacies studies in thin sections, aiming at the documentation of the lateral
Figure 2.4 Well locations and area subdivision of Bass Islands Formation and Bertie Formation in SW Ontario. Area 1, Area 2 and Area 3 are divided by distinct lithofacies motif. Area 1 belongs to the eastern margin of Michigan Basin. Area 2 locates at southeastern margin of Michigan Basin and is adjacent to northern Ohio. Area 3 along the western shoreline of Lake Erie are adjacent to the Appalachian Basin in units in western New York. Spaces between areas are lacking cored well data.
and vertical lithologic changes, defining distinct formational or lithofacies contacts, interpreting the depositional history of the targeted strata and exploring the geologic controls on regional bedrock aquifers formation and distribution. Photographs of key cores and outcrops are used to support interpretation of each lithofacies (Fig 2.5-2.12). The carbonate lithofacies classification and porosity geometry were described using Dunham’s (1962) and Choquette and Pray’s (1970) classification schemes.

The underlying Salina G Unit consists of a cycle of lower evaporite-bearing microbialites and upper dark grey shale with anhydrite nodules. The occurrence of dark grey, dolomitic shale and the disappearance of evaporites define the basal contact of Bass Islands Formation and Salina Group in cored wells. The stratigraphic positions and thickness of the common peritidal and sabkha lithofacies within Bass Islands and Bertie formations vary locally.

In Area 1 and 2, the Bass Islands Formation comprises thicker and more restricted evaporite-carbonate lithofacies than those in Area 3. The Bass Islands Formation in area 2 is complete, overlain by Lower Devonian carbonates that is considered to be equivalent to the Rondout and Manilius Formations in New York (Brett et al. 2000).

Major facies patterns have been recognized in all three areas. Area 1 and 2 have microbial laminites, thrombolites, argillaceous dolomudstone, evaporites, subaerial brecciations and karstic dissolution structures. In Area 3, evaporites are absent and the dolostone matrix is less argillaceous. Karstic features are common in all three areas, at both lower and top parts of the Bass Islands Formation. Weathering and meteoritic dissolution structures are very common in the Bass Islands Formation (Brunton and Dodge 2008).

In Area 3, though the Bertie Formation has long been regarded as equivalent to the Bass Islands Formation in area 1 and 2 (Sanford 1969; Johnson et al. 1992), recent studies by Haynes
(1991) and Haynes and Parkins (1992) showed a possible alternative correlation of the Bass Islands Formation and Bertie Formation. Based on lithology and depositional pattern of the Bertie Formation in both outcrops and subsurface, Haynes (1991) has observed that the Akron Member at the top of Bertie Formation in Ontario is overlain by the brecciated dolostones in the Bass Islands Formation. This suggests that the Akron Member should be either correlated to the Salina G and F unit in Michigan and Ohio. Alternatively, this contact relationship can be interpreted as the results of the initial Bertie sea transgressing from Ohio rather than from Michigan, which later retreated to New York, and then inundated by the Bass Islands sea from Michigan.

2.4 Lithofacies of the Salina G Unit and the Bass Islands Formation

Lithofacies description and distribution of the Salina G Unit and the Bass Islands Formation are listed in Table 2.1 based on lithologic features and sedimentary/diagenetic structures. In general, 15 lithofacies have been recognized and integrated with geophysics and fauna studies to provide lithological information for further studies of paleoecology and regional stratigraphic correlation. Some facies within the Bass Islands Formation are overprinted by diagenetic features that contribute to the interpretation of the post-depositional environment of the lithofacies, including replacement, karstic events, subaerial dissolution and deep burial dissolution.

In area 1 and 2 in the Michigan Basin area, a “False G” unit is present in the middle of the Bass Islands Formation (Armstrong and Carter 2010). It consists of interbeds of anhydrite nodules and dark grey to medium brown, microbial laminated to bluish grey mottled dolostones. Though it resembles the Salina G Unit in lithology, its stratigraphic position and distinctive mottled patterns are readily distinguishable from the Salina G Unit. It represents the shallowest deposits in the Bass Islands Formation and corresponds to a high excursion in the gamma ray log because of high
contents of clay minerals as undissolved residues. This high response can be regionally correlated to the argillaceous bed that separates the Akron Member and the overlying Bass Islands Formation in area 3 within the Appalachian Basin. Multiple karstic dissolution of evaporite intervals were present in the “False G” unit to form dissolution-collapse structures.
Table 2.1 Lithofacies types of the Salina Group G Unit (SG-1 to SG-3) and the Bass Islands Formation (BI-1 to BI-11). BI-12 possibly represents the lithofacies of the Rondout Formation described by Brett et al. (2000) in New York. (see Figures 2.1 for stratigraphy, and Figure 2.5-2.12 for core photos).

<table>
<thead>
<tr>
<th>Lithofacies (LF) Description</th>
<th>Thickness, Nature of Contacts and Distribution</th>
<th>Depositional Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>SG-1. Argillaceous Dolomudstone with Anhydrite Laths.</strong> Buff to tan, thinly or nodular bedded dolomudstone with intraclastic intercalations and rip-up clasts (Fig 2.5-1 and 2.5-2). Wispy argillaceous seams are very common. Various types of gyspums or anhydrite crystals or molds after gypsum are present in different shapes: irregular, rosette-like or needle-like. Evaporite clusters tend to accumulate along bedding planes or along argillaceous seams, replacing previous plastic sediments. Locally, 0.5-1 mm thick anhydrite flakes are observed formed by coalesced anhydrite minerals.</td>
<td>2-3.5 m thick. Lower Salina G in Area 1 and 2. Absent from the Appalachian Basin, where the Salina F greenish grey shale is overlain by SG-2 with an undulatory dissolution surface. The basal contact with Salina F Unit is marked by a 50-cm thick, dissolution-induced brecciated interval with common subangular clasts and dark brown dolomuds infilling the open space.</td>
<td>Subtidal. Crystals grew under constant water cover; modern analogue is Gulf of Mexico. (Warren 2016)</td>
</tr>
<tr>
<td><strong>SG-2. Microbial-laminated Dolomudstone with Anhydrite.</strong> Medium to dark greenish grey or brown, thinly laminated, microbial to stromatolitic dolomudstone (Fig 2.5-2 to 2.5-4). Pisolitic fabrics are concentrated in up-section, replaced by anhydrite minerals. Lamination are 0.2-0.5 mm layers of fine- to microcrystalline-dolostone, with common pisoliths, intraclasts, mud-cracks, tepee structures, and degassing or dewatering pipes. Dolostone layers are locally distorted by bluish grey, irregularly folded, replacive anhydrite nodules. Contact of anhydrites and dolostone matrix are diffused. Incipient anhydrite nodules are locally present as less than 2 cm in size, micro-aggregates or lathlike, after gypsum, which in turn replaced the original dolomite. Up-section anhydrites are mosaïc, chicken-wire, enterolithic and massive, and microbial-laminate fragments irregularly filled the inter-nodular space. Coalescing nodules cap exhibit massive chicken-wire fabrics with micro-nodular anhydrites cohesive by clayey carbonate muds.</td>
<td>Lower Salina G in Area 3. Absent from the Michigan Basin area, where the SG-1 overlies the Salina F greenish grey shale with anhydrite nodules directly. In Appalachian area, this lithofacies overlies the Oatka Member of the Bertie Formation and is equivalent to lower Falkirk Member. The basal contact is marked by a dissolved horizon or anhydrite laths rich argillaceous seams. Upper contact with SG-3 is gradational, marked by a change of anhydrite forms from dispersed laths to large replacive nodules.</td>
<td>Intertidal to lower Sabkha, with periodic subaerial exposures and rapid fluctuation of sea level in a very shallow, restricted hypersaline environment.</td>
</tr>
</tbody>
</table>
**SG-3. Dolomitic Shale with Anhydrite.** Dark grey, thinly to medium bedded, dolomitic shale with light grey to white anhydrites (Fig 2.5-5). Orange anhydrite nodules are locally developed due to iron staining. Anhydrites are present in two forms: replacive nodular and mosaic-like. The displacive mosaic-like anhydrites show typical vertical growth layers that distort the dolomitic shale matrix. Replacive anhydrite nodules usually underlie the mosaic-like anhydrites, surrounded by argillaceous seams, resembling their dolomite precursors. Anhydrite flakes are locally present as 0.5-1 cm thick horizontal beds.

**BI-1. Dolomitic Shale.** 0.3-1.5 m thick, dark grey, thinly bedded dolomitic shale or clayey dolomudstone above the contact of Bass Islands Formation and Salina G Unit (Fig 2.6-1 to 2.6-5). Within 30 cm above the basal contact, concentrated carbonaceous seams are well developed with anhydrite detritus less than 1 mm in diameter, and dolomitic shale beddings are partly contorted or brecciated. Clasts are confined by irregular argillaceous seams. Color grades into medium grey in up-section, with less clay content, and beddings become wavy to normal.
BI-2. **Thrombolitic Dolomudstone-Dolowackestone.** Light to medium brown, poorly bedded, finely crystalline or recrystallized, bioturbated dolomudstone to wackestone with distinctive moldic porosity and irregular pisolitic vugs (Fig 2.7-1 to 2.7-5). Dolomudstone matrix are commonly recrystallized and show clotted fabrics with irregular black, carbonaceous stylolites. Brecciated zones are commonly developed and are surrounded by argillaceous seams. The open space among breccias are partly plugged by secondary anhydrites or dolomites. Brownish grey, fenestral mottles impart the clotted fabrics due to gastropod (?) bioturbation, but megafossils are lacking. Anhydrite laths are uncommon and are commonly replaced by dolomite.

In the Appalachian Basin area, the 1-3.5 m thick BI-2 overlies the clayey dolostone at the top of BI-1 by a sharp contact or an inclined karstic surface. The sharp contact is marked by a 2-mm thick dark grey wispy argillaceous seam. The karstic interval is 30-35° inclined, with common poorly sorted small autogenetic breccias and iron stains. In the Michigan basin area, BI-2 is absent in lower Bass Islands and occurs in upper Bass Islands above “False G” unit. It overlies the evaporitic microbialites by a sharp lithologic change from thinly bedded algal laminites of “False G” into its basal poorly bedded thrombolitic dolostone.

In the shallow subtidal to lower intertidal Bass Islands Formation in area 1, 2 and 3.

**BI-3. Microbial-laminated Dolomudstone.** Two subtypes: BI-3a. Thinly laminated, bluish grey, cryptocrystalline microbial-laminated dolomudstone (Fig 2.8-1 to 2.8-5). Microbial laminae are 1-2 mm thick, undulating or distorted; BI-3b. Medium to dark brown, thinly bedded, microbial-laminated dolomudstone with intraclastic interbeds (Fig 2.8-6 to 2.8-9). Laminated beddings are 1-3 cm thick, commonly exhibiting planar or convoluted fabrics. Intraclasts are light tan to buff, autogenetic mudstone pebbles, less than 5 mm in diameter, well-rounded, floating within dolostone mud matrix. Megafossils are very rare, but fragments of dark grey shells are found. Both subtypes contain subaerial exposure structures as desiccation cracks, flat-pebbles, tepee structures and bluish grey fenestral mottles. Crack infills and caliche horizons are locally present. Stromatolites fabrics are observed in an isolated interval below the BI-3a subtype in Core 1. U.S Steel No.1.

This lithofacies prevails in Bass Islands Formation and occurs at various intervals. The thickness varies from less than 50 cm to over 5 m. It commonly overlies the BI-2 by a gradational lithologic change. Karstic surfaces are usually well developed on top of BI-3.

In the shallow subtidal to lower intertidal Bass Islands Formation, in area 1, 2 and 3.

Supratidal to lower Sabkha.
**BI-4. Brecciated Microbialites and/or Dolomudstone:** Two subtypes: BI-4a. Brecciated microbial dololaminite. Light to medium brown or bluish grey, clast-supported brecciated interval with very common dissolution-collapse structures (Fig 2.9-1 to 2.9-4). Clasts are angular or subangular, microbial laminated dolostone in a grey to brown recrystallized matrix. Clast size ranges from less than 1 mm to 5 cm. Clasts decrease in size up-section, from lower rectangular to horizontally elongate form to upper gently undulated and distorted irregular laminate form. Dissolution joints commonly truncate the microbial laminties, with small dolostone pebbles and recrystallized dolomudstone infilling the joints. Locally, caliche textures are present resembling this lithofacies subtype. BI-4b. Brecciated Dolomudstone. Clast or matrix supported, light tan to bluish grey, thin persistent karstic facies (Fig 2.9-5 to 2.9-9). The base comprises clast-supported, medium brown, pelloital dolomudstone breccias with recrystalline matrix. Up-section, mudstone clasts are highly distorted and supported by bluish grey recrystalline matrix. The clotted matrix fabrics resemble anhydrite nodules after gypsum, indicating an anhydrite-dissolution nature of the breccias.

**BI-5. Bioturbated Dolomudstone.** Medium to massive bedded, light tan to grey, microcrystalline dolomudstone with bluish bioturbation fabrics (Fig 2.10-1 to 2.10-4). Fossils are uncommon but include bivalve shells and gastropods. Patches of moldic porosity are created by dissolution of gastropod or bivalve(? ) shells. The intensive bioturbation obliterates the bedding planes. Bioturbation is bluish grey to buff in color, irregular and diffused into the brownish grey dolomudstone matrix. Peaked argillaceous seams are commonly present in Michigan area.

Bass Islands Formation in area 1, 2 and 3. BI-4 represents a dissolution induced karstic lithofacies, and the thickness depends on the extension of karstic systems. Usually it is 50-80 cm thick where it overlies BI-3, but reaches more than 1m where it overlies the crackly fractured dolomudstone of BI-6. In both cases, brecciation cracks, karstic dissolution horizons, carbonate crusts, caliches and secondary evaporite minerals occur in the uppermost of the underlying facies. BI-4 is abruptly terminated by karstic dissolution at the top with a sharp lithologic change with the overlying units.

Lower Bass Islands Formation in area 2 and 3. Thickness ranges from 50-80 cm, commonly overlies BI-3 by a gradational lithologic change or overlies a karstic surface directly. Usually underlies a brecciated zone with dissolution collapse structure.

**Supratidal to lower Sabkha**

**Shallow subtidal**
BI-6. **Homogeneous Dolomudstone.** Buff to tan, very fine to microcrystalline, commonly leached or recrystallized dolomudstone (Fig 2.10-5 to 2.10-8). Shell debris are uncommon. Crackle fractures and fissures are well developed in some intervals within this lithofacies (Fig 2.10-5). Fissures and joints are either vertical or horizontal, truncating the dolomudstone matrix to form a clast-supported interval. The clasts are composed of angular, poorly sorted homogeneous dolomudstone, with size ranging from 5-20 mm. Fissures and joints are commonly infilled by dark grey dolomitic muds, but can also be cemented by whitish grey secondary dolomites. In Michigan area, this lithofacies is more argillaceous, with common peaked stylolites. Burrows are found capping this lithofacies and infilled by overlying microbial laminites (Fig 2.10-6). Gypsum clusters or needle-like gypsum laths are commonly found above the bedding planes (Fig 2.10-7 and 2.10-8).

BI-7. **Porous Dolomudstone with Evaporites.** Thin to medium bedded, faintly laminated, bituminous dolomudstone with very common fenestral vugs that are partly plugged by anhydrites (Fig 2.11-1 and 2.11-2). The dolomudstone matrix contains common discontinuous carbonaceous streaks. Fenestral vugs are parallel to the beddings. Degassing or dewatering structures are common. The growth and dissolution of anhydrite infills have increased the vug size.

Bass Islands. Area 1, 2 and 3. In Appalachian Basin areas, BI-6 ranges from 1.2-3.5 m, overlying a karstic surface with steep relief; by irregular karstic dissolution horizon or on undissolved shale residues of cave infills. The basal contact usually accompanies a distinct lithologic change of underlying brecciated microbialites or dolomudstone to the overlying thinly laminated microbial dolostone. In Michigan Basin area, BI-6 are less fractured and the basal contact may be marked by a gradational lithologic change from laminated microbial dolostone or by a burrowed surface.

Restricted, low-energy, subtidal environment of relatively shallow water depth.

Bass Islands in area 2 and 3. Thickness is 20-30 cm. Overlies the brecciated microbialites of BI-4 directly.

Lower intertidal to restricted, low energy shallow lagoon. Crystals grow under constant water cover in early diagenesis.
**BI-8. Argillaceous Dolomudstone.** Medium to massive bedded, light to medium grey, argillaceous, locally sucrosic dolomudstone with pervasive vugs (Fig 2.11-6 to 2.11-8). Wispy argillaceous seams are very common, and vugs are highlighted along the seams by groundwater dissolution. Vertical fissures are well developed and are partially plugged by anhydrite minerals. General leaching of fresh water hindered the observation of dolomudstone fabrics. The unaltered lithofacies may be analogous to BI-6 in Michigan area.

Bass Islands Formation in area 1. Restricted to Michigan Basin area. Thickness ranges from 0.3-4.5 m. It either underlies the evaporitic “False G Unit” in middle Bass Islands Formation or presents in various interval in upper Bass Islands Formation. Contact with the underlying facies are seen with a gradational lithologic change by concentration of wispy stylo-seams. BI-8 either grades upward into dolomudstone with anhydrite nodules of BI-9 or capped by a leaching interval with dissolution highlighted porosity.

**BI-9 Mottled Dolomudstone with Anhydrite.** Medium to dark grey, medium bedded, cryptocrystalline and locally clayey dolomudstone with characteristic irregular and discontinuous dark grey mottles (Fig 2.12-1 and 2.12-6). Mottles are variable in size, generally less than 2 cm, diffused contact with the dolomudstone matrix. The similar fabrics within mottles and dolomudstone matrix in thin sections indicate they are not burrow traces except for a change of organic contents. Clay minerals are locally present, thus increasing the gamma ray value response. Anhydrite flakes or nodules are recognized as replacing or capping the mottled dolomudstone. Anhydrites are light to bluish grey, mosaic-like with common bluish seams surrounding each nodule. The surrounding dark grey seams are high in clayey contents, indicating an incipient form of the dark grey mottles.

“False G Unit” in Bass Islands Formation in area 1 and 2. In the middle Bass Islands Formation, the BI-9 overlies the mottled dolomudstone resembling the anhydrite precursors. The contact is gradational. The mottled dolomudstone grades upwards into interbedded anhydrite nodules and dolomudstones. The basal contact is chosen at the first appearance of anhydrite bed. This lithofacies can be easily picked up by high response of gamma ray log.

“Shallow subtidal to lower intertidal.”

**Very shallow water and lower sabkha; minor fluctuation of sea level on a shallow ramp.**
BI-10. Microbial-laminated Dolomudstone with Anhydrite. Two subtypes. BI-10a is the same as SG-2 (Fig 2.12-2 and 2.12-3). The only difference is that in the dolomudstone matrix of BI-10a the anhydrite interlayers are blue in color and thicker (1-6 cm thick). BI10b. Buff to tan, thinly bedded microbial or stromatolitic dolomudstone with very common anhydrite nodules and laths (Fig 2.12-5 to 2.12-7). Dolomudstone beds are 1-2 cm thick. Microbially laminated layers contain small dolomitic flat-pebbles, desiccation cracks. Anhydrite nodules are light grey, partly iron stained, displacive, showing clotted fabrics with argillaceous seams among each nodule. Dolostone beddings are commonly distorted by the overgrowth of anhydrite nodules converted from gypsum. Anhydrite laths are brown to brownish grey, parallel to beddings planes and accumulate around interlaminations.

BI-11. Bituminous Gastropod/Ostracod-Pelletal Dolowackestone. Thinly to medium, partly pseudonodular bedded, medium to dark brown, bituminous dolowackestone with common gastropod moldic porosity and pellets (Fig 2.11-3). Beddings are pseudonodular, with common bituminous interbeds. Dolomudstone is highly bioturbated and burrow traces are common along bedding surface. Fragments of gastropods, bivalves and ostracods are locally abundant. Moldic porosity of ostracods are common in area 2. Concentrated carbonaceous seams are developed. Karstic surfaces characterized by iron stains and 1-2 cm undulatory beds of undissolved residues are recognized in up-section.

“False G Unit” in Bass Islands Formation in area 1 and 2. BI-10 overlies BI-9 or BI-3. Where it overlies BI-9, the contact is sharp and undulatory, marked by concentrated stylo-seams. Where it overlies the microbial laminites in of BI-3, the contact is gradational. The microbial laminae of BI-3 thins up-section and the first occurrence of anhydrite nodule marks the contact. BI-9 and BI-10 altogether are 10-12 m thick.

BI-11. Bituminous Gastropod/Ostracod-Pelletal Dolowackestone. Thinly to medium, partly pseudonodular bedded, medium to dark brown, bituminous dolowackestone with common gastropod moldic porosity and pellets (Fig 2.11-3). Beddings are pseudonodular, with common bituminous interbeds. Dolomudstone is highly bioturbated and burrow traces are common along bedding surface. Fragments of gastropods, bivalves and ostracods are locally abundant. Moldic porosity of ostracods are common in area 2. Concentrated carbonaceous seams are developed. Karstic surfaces characterized by iron stains and 1-2 cm undulatory beds of undissolved residues are recognized in up-section.

Upper Bass Islands in area 2 and 3. Restricted to Appalachian Basin area. Thickness ranges from 1.8-2.5m. BI-11 Usually underlies the karstic interval below S-D contact and overlies the lower leached unit. The lower contact is marked by a sharp peaked stylolite that represent a dissolution surface. Uppermost BI-11 is capped by a bluish mottled unit that marks the base of karstic interval below contact.
| **BI-12. Ostracod-Gastropod Dolowackestone.** Cream to light tan, massive bedded, bioturbated, very porous, pelloidal dolowackestone with abundant ostracods and gastropods moldic pores (Fig 2.11-4 and 2.11-5). Ostracods and gastropods are abundant and all dissolved, leaving the moldic porosity partly plugged by secondary dolomites. The dolostone matrix is highly bioturbated, showing a mottled texture. Argillaceous seams are very rare. | Possible equivalent to the Rondout-Manilius Formation in area 2. This lithofacies is restricted to two cores (Consumers’ 33409 and Cansalt DDH 87-3). It is present at the top of Bass Islands Formation in area of western Lake Erie shoreline that can possible be correlated to lower Rondout Formation in New York. It overlies the BI-3a by a sharp lithological change from restricted peritidal microbial-laminated dolomudstone to fossiliferous dolowackestone. In core Consumers’ 33409, this lithofacies is overlain by dark grey, stromatoporoid-coral floatstone that can be recognized as upper Rondout Formation-Manilius Formation. | Subtidal to shallow continental shelf. |
**Figure 2.5:** Lithofacies variations in Salina G Unit.

1. Lithofacies SG-1 (Argillaceous Dolomudstone with Anhydrite Laths) in area 3. Buff, finely crystalline, thinly bedded dolomudstone with very common wispy argillaceous stylo-seams (Core 3. Consumers Pan Am 13058). Small dolomite pebbles tend to accumulate within argillaceous intervals formed by scouring or dissolution. Anhydrite laths are euhedral, brownish grey, sparsely dispersed in dolomudstone matrix, representing diagenetic features.

2. Lithofacies SG-2 (Microbial-laminated Dolomudstone with Anhydrite) and SG-3 (Dolomitic Shale with Anhydrite) in area 3 (Core 2. U.S Steel. No.1). SG-2 (Left): Greenish grey, thinly bedded, microbial-laminated dolomudstone with abundant anhydrite nodules. Growth of anhydrite nodules have distorted the microbial growth lines. SG-3 (Right): Dark grey, thinly to medium bedded dolomitic shale with light grey anhydrite nodules. Note the dissolution-collapse structure occurs overlying the large displacive chert nodule in the middle.

3. Contact of Salina G Unit and Bass Islands Formation in area 1 (Core 1. DOMTAR GODERICH S.T.#1). Lower Salina G Unit consists of dark grey dolomitic shale with anhydrite flakes. The disappearance of anhydrite marks the contact of Salina Group and Bass Islands Formation. Basal Bass Islands Formation is featured as massive bedded, dolomitic shale.

4. Lithofacies SG-3 (Dolomitic Shale with Anhydrite) in area 2 (Core 2. U.S Steel. No.1). The upper dolostone drapes the lower anhydritic shales by a sharp and irregular surface. Below the surface, the dolomitic shale contains incipient replacive anhydrite minerals. Up-section, the dolostone beddings are irregular and locally truncated, representing various dissolution surfaces.

5. One cycle of SG-1, SG-2 and SG-3 in area 3 (Core 4. Consumers' Amoco 13061). The mosaic-like anhydrite nodules cap this shallowing cycle, representing the shallowest sabkha deposits. Up-section is SG-3, anhydrite nodules show typical vertical growth pattern that are interpreted as subaqueous evaporites.
Figure 2.6: Lithofacies of BI-1 (Dolomitic Shale) in the Bass Islands Formation.

1. Contact of the Salina G Unit and the Bass Islands Formation in area 3 (Core 2. U.S Steel. No.1). The contact is marked by an undulatory surface that is overlain by concentrated argillaceous seams with light grey clasts floating within. Basal Bass Islands Formation shows chaotic fabrics with irregular dolomitic shale breccias surrounded by shaly seams. These irregular or nodular beddings were formed by partly dissolution of the dolomitic materials.


3. Contact of BI-1 and BI-2 in area 3 (Core 9. Consumers' Amoco 13076). The contact is marked by an inclined surface with a dissolution joint extended into the lower BI-1. Anhydrite flakes is 1.5 cm above the contact. The chaotic fabrics in upper BI-2 represent various dissolution surfaces.

4. Contact of SG-3 and BI-1 in area 3 (Core 11. OGS 82-3). The concentrated shale seams with dolomite pebbles overly the anhydrite nodules growing within microbial-laminated dolomudstone by an irregular surface.

5. Dolomitic shale in area 2 (Core 19. Consumers 33409). Dark grey, thinly bedded dolomitic shale with common carbonaceous seams. Carbonaceous seams are concentrated in the lower part. Pyrites are present in the middle.
**Figure 2.7:** Lithofacies of BI-2 (Thrombolitic Dolomudstone).

1. BI-2 in area 1 (Core 8. Imperial No. 809 - W.J. Mawson No. 1). Massive, greenish tan to grey mottled dolomudstone with common irregular vugs and fissures. Dolomudstone matrix is highly recrystallized. Wispy stylolites are disorderly arranged due to the growth of replacive anhydrites.

2. BI-2 in area 2 (Core 16. OGS 82-2). Similar to BI-2 in area 1, massive bedded, greenish tan to grey mottled, recrystallized dolomudstone. Fenestral vugs are plugged by anhydrites. Argillaceous stylolites are discontinuous.

3. BI-3b (left) and BI-2 (right) in area 3 (Core 2. U.S. Steel No.1 - J. H. Lawrence No. 1). Left: BI-3b overlies the subtidal dolowackestone with sparsely bivalve shell fragments by a sharp contact of anhydrite flake. A solution-widened joint is present below the contact and is plugged by anhydrites. Right: Massive bedded, tan pelloital dolomudstone with anhydrite laths.

4. BI-2 in area 3 (Core 10. Consumers' Pan Am 13057). Brecciated, tan to light grey, thrombolitic dolomudstone with very common dissolution-induced cavities that have been plugged by secondary dolomites.

5. BI-2 in area 2 (Core 22. Cansalt DDH 87-3). Greenish grey, massive bedded, recrystallized dolomudstone with very common irregular vugs. Dissolution pipes are present and lower part of which are infilled by secondary dolomite.
Figure 2.8: Lithofacies of BI-3a and BI-3b (Microbial-laminated Dolomudstone).

1. BI-3a in area 3 (Core 2. U.S Steel. No.1). Light grey, algal-laminated dolomudstone. Laminae are partly distorted by the sparse dark bluish grey mottles. A dissolution surface that is covered by anhydrite interbeds is present in the middle, and is overlain by subsequent cross-beddings. Blue mottles are elongate along bedding planes, indicating a decay or degassing origin of the microbial mats.

2. Contacts of BI-3a and BI-4a in area 3 (Core 2. U.S Steel. No.1). The upper laminated dolomudstone overlies the brecciated microbial-laminated dolomudstone by irregular stylolites. The lithofacies of breccias are BI-3a, and the beddings are distorted by dissolution and collapse. Uppermost brecciated interval is anhydritic.

3. Interbeds of BI-3a and BI-4a in area 3 (Core 2. U.S Steel. No.1). Microbial laminites are locally crinkly and distorted. Bluish grey mottles are in two forms: (1) the small, oval to elongate mottles that are parallel to bedding planes, and (2) the large irregular blueish grey mottles diffused in microbial-laminated dolostone. The later form shows a peneadiagenetic discoloration process by groundwater flow.

4. BI-3a in area 3 (Core 3. Consumers' Pan Am 13058) with common fenestral vugs in the upper section. Laminae are locally distorted.

5. BI-3a in area 3 (Core 4. Consumers' Amoco 13061). Pyritic, crinkly laminated, microbial dolomudstone overlies the brecciated dolomudstone of BI-4a by a 2-mm thick anhydrite flake.

6. BI-3b in area 3 (Core 10. Consumers' Pan Am 13057). Light tan to brown, thinly bedded dolomudstone with common fractures and concentrated carbonaceous stylo-seams.

7. BI-3b in area 3 (Core 11. Consumers' Pan Am 13057). Light bluish brown, thinly and irregularly bedded dolomudstone with common desiccation cracks.

8. BI-3a in area 2 (Core 16. OGS 82-2). Light grey, thinly and crinkly laminated microbial dolomudstone with teepee structure and fenestral vugs. Bluish grey mottles are pervasive.

9. BI-3a in area 2 (Cansalt DDH 87-3). Light tan to grey, thinly laminated, argillaceous dolomudstone with desiccation cracks in the lower section.
Figure 2.9: Lithofacies of BI-4a and BI-4b (Brecciated Dolo-microbialites-Dolomudstone).

1. BI-4a in area 1 (Core 1. DOMTAR GODERICH S.T.#1). Two brecciated zones interbedded with microbial-laminated dolomudstone. The lower brecciated zone is 5 cm thick, within which the laminae are contorted by dissolution of impurity-rich interval. Upper brecciated zone is 4-5 cm thick, within which subangular microbialite clasts floating in the pisolitic dolostone matrix.

2. BI-4a in area 3 (Core 2. U.S Steel. No.1). The lower microbial laminites are distorted by meteoric dissolution, showing caliche-like fabrics, which is overlain by blue mottled, chaotic dolomudstone with common fenestral vugs.


4. BI-4a in area 3 (Core 4. Consumers' Amoco 13061) Brecciated microbialite clasts floating in a cryptocrystalline matrix. The matrix is recrystalline, replacing after the anhydrite precursors.

5. BI-4b in area 1 (Core 8. Imperial No. 809 - W.J. Mawson No. 1). Tan to light grey, thinly bedded dolomudstone with two intraclastic intercalations. Rip-up clasts and scouring surface are present, indicating a possible periodic subaerial exposure of the dolomudstone.

6. BI-4b in area 1 (Core 13. ARGOR 65-1). Caliche fabrics in the dolomudstone matrix with the lower solution-widened joint.

7. BI-4b in area 2 (Core 16. OGS 82-2). Buff, brecciated, dolomudstone with bluish grey anhydritic matrix. Anhydrites are commonly replaced by dolomites.

8. BI-4a in area 2 (Core 22. Cansalt DDH 87-3). Light grey, brecciated, microbial-laminated dolomudstone. Solution joint connects the subaerial surface and the lower meteoric percolation zone. In dissolution cavities, microbialite clasts lie on the cavity floor and is capped by fine-grained cavity infills. Concentrated carbonaceous seams overlie the brecciated unit by a sharp flat surface.

9. BI-4b in area 2 (Core 22. Cansalt DDH 87-3). Light grey to cream, brecciated and argillaceous dolomustone with common low-relief peaked stylolites.
Figure 2.10: Lithofacies of BI-5 (Bioturbated Dolomudstone) and BI-6 (Homogeneous Dolomudstone).

1. BI-5 in area 1 (Core 1. DOMTAR GODERICH S.T.#1). Bluish grey, massive bedded, homogeneous dolomudstone with bluish grey mottles and peaked stylolites. Fossils are very rare.
2. BI-5 in area 3 (Core 2. U.S Steel. No.1). Light brown to brownish grey, massive bedded, bioturbated dolomudstone. Fragments of gastropods are present in the up-section.
3. Contact of BI-3a and BI-5. (Core 3. Consumers' Pan Am 13058). The light tan, pelloital, bioturbated dolomudstone-wackestone overlies the thinly bedded, microbial-laminated dolomudstone by an irregular dissolution surface. Blue mottles and bivalve fragments are present in the overlying BI-5.
4. BI-5 in area 2 (Core 4. Consumers' Amoco 13061) Greenish grey, massive bedded, mottled dolomudstone with pyrite concentrations along bedding planes, indicating a reducing environment.
5. BI-6 in area 1 (Core 1. DOMTAR GODERICH S.T.#1). Creamy to buff, massive bedded, pelloital dolomudstone with very common crackle fractures and brecciated intervals.
6. Contact of BI-5 and BI-6 in area 3 (Core 4. Consumers' Amoco 13061). A burrowed surface of the underlying greenish grey homogeneous dolomudstone is overlain by the bioturbated dolomudstone of BI-5 with common carbonaceous partings.
7. BI-6 in area 3 (Core 10. Consumers' Pan Am 13057). Light grey, massive bedded, pelloital dolomudstone with rare shell fragments. Diagenetic gypsum cluster is present in the lower section.
8. BI-6 in area 3 (Core 11. OGS 82-3). Light grey, massive bedded, pelloital dolomudstone with faint colored dissolution joints. Secondary needle-like gypsum casts are abundant and dispersed in the pelloital dolomudstone.
**Figure 2.11:** Lithofacies of BI-7 (Porous Dolomudstone), BI-8 (Argillaceous Dolomudstone), BI-11 (Bituminous Gastropod-Pelloital Dolowackestone) and BI-12 (Ostracod-Gastropod Dolowackestone).

1. BI-7 in area 3 (**Core 2. U.S Steel. No.1**). Light greenish grey, faintly laminated dolomudstone with blue mottles and fenestral vugs. Vugs are partly plugged by secondary dolomite.
2. BI-7 in area 3 (**Core 3. Consumers' Pan Am 13058**). Greenish grey, faintly laminated dolomudstone with very common fenestral vugs that are partly plugged by dolomites. Growth rings of dolomites are recognizable. Microbial-laminated dolomudstone is present in up-section.
3. BI-11 in area 3 (**Core 2. U.S Steel. No.1**). Medium brown, thin-medium bedded, bituminous dolowackestone with common dark grey shell fragments.
4. BI-12 in area 2 (**Core 19. Consumers 33409**) Greenish grey, massive bedded, gastropod and ostracod rich dolowackestone. Abundant moldic pores of gastropods and ostracods are present. Possible Rondout Formation.
5. BI-12 in area 2 (**Core 22. Cansalt DDH**). Greenish grey dolowackestone with moldic porosity of ostracods. The dolostone matrix is partly recrystallized. Possible Rondout Formation.
6. BI-8 in area 1 (**Core 1. DOMTAR GODERICH S.T.#1**). Light grey, massive bedded, argillaceous dolomudstone with common vugs parallel to bedding planes. Peaked stylolites are present.
7. BI-8 in area 1 (**Core 8. Imperial No. 809 - W.J. Mawson No. 1**). Massive bedded dolomudstone with moldic gypsum clusters developed at dissolution surfaces. Gypsum was dissolved after crystal formation and the mold remained unplugged.
8. BI-8 in area 2 (**Core 22. Cansalt DDH 87-3**). Light grey, massive bedded, pelloital dolomudstone with argillaceous seams and vugs at the bottom.
Figure 2.12: Lithofacies of BI-9 (Mottled Dolomudstone with Anhydrite) and BI-10 (Microbial-laminated Dolomudstone with Anhydrite).

1. BI-9 in area 1 (Core 1. DOMTAR GODERICH S.T.#1). Left: Medium grey dolomudstone with bluish dark grey mottles, resulting from the growth of anhydrite nodules and subsequent replacement of the dolomite. Right: Microbial-laminated dolomudstone with bluish grey, irregular anhydrite nodules. Anhydrite nodules grew within the microbial laminites and replace part of the dolomite matrix.

2. BI-10 in area 1 (Core 1. DOMTAR GODERICH S.T.#1) Light grey, mosaic-like anhydrite nodules surrounded by argillaceous seams is overlain by concentrated stylolites.

3. BI-10 in area 1 (Core 8. Imperial No. 809 - W.J. Mawson No. 1). Left: Light grey, microbial-laminated dolomudstone with replacive anhydrites and common anhydrite laths. Laminae are contorted as a result of anhydrite growth. Right: A shallowing-upward cycle consists of lower microbial-laminated dolomudstone with replacive anhydrites and an upper light grey anhydrite nodules on top.

4. Contact of BI-9 and BI-10 in area 1 (Core 12. Imperial 805 – Lyons No.1). The contact of lower mottled dolomudstone and upper dolomudstone with blue anhydrite nodules is gradational, marked by the first appearance of replacive anhydrite.

5. BI-10 in area 1 (Core 12. Imperial 805 – Lyons No.1). The greenish grey microbial-laminated dolomudstone and the blue anhydrite nodules have a sharp or diffused contact.

6. BI-9 in area 1 (Core 13. Argor 65-1). Light grey dolomudstone with bluish grey, irregular mottles that is overlain by replacive anhydrite nodules. Pervasive iron stains indicate an alteration by groundwater in diagenesis.

7. BI-10 in area 2 (Core 16. OGS 82-3). Thinly laminated microbial dolomudstone with anhydrite nodules interbeds. The growth of anhydrites has distorted the original laminae.

8. BI-10 in area 2 (Core 22. Cansalt DDH 87-3). Massive dolomudstone that are partly dissolved, showing chaotic fabrics and the dissolution cavity is plugged by secondary anhydrite nodules.
2.5 Facies Groups and Regional Distribution

The lithofacies descriptions and general interpretations have been provided in Table 3.1, with their thickness, regional distribution and nature of contacts briefly discussed. The Bass Islands Formation contains 12 lithofacies (BI-1 to BI-12) and Salina G Unit contains 3 (SG-1 to SG-3). The 15 lithofacies have been assigned into 7 Facies Groups, showing general shoaling depositional patterns within each group. Some lithofacies occur in multiple Facies Groups.

2.5.1 Shallow Subtidal Facies (Facies Group 1)

Facies Group 1 contains lithofacies BI-2, BI-5, BI-6 and BI-8. This Facies Group contains various end members that were formed in shallow subtidal, oxygenated water realm above the normal wave base level, with periodically leaching and dissolution by meteoric water in later diagenetic processes. Lithofacies vary due to the presence of bioturbation, argillaceous materials content and dolostone heterogeneity.

BI-2 occurs throughout Area 1, 2 and 3, overlying the basal dolomitic shale (BI-1) of the Bass Islands Formation and reappears above the “False G Unit”. Most commonly this thrombolitic dolomudstone exhibits strong diagenetic overprints, including dissolution pipes and recrystallized dolostone matrix (Fig 2.7). Cavities are commonly found plugged by secondary dolomite sparite (Fig 2.7-4). In lower BI-2, faintly laminated and randomly oriented laminae are dispersed in the thrombolitic dolostone matrix with common dissolution collapse structure (Fig 2.7-1). The collapse clasts are subangular, 5–10 cm in size, surrounded by argillaceous or carbonaceous stylolites. Up-section, BI-2 either grades into the microbial-laminated dolomudstone (BI-3) or warped by anhydrite nodules at the top. On a specific level, the basal thrombolitic mudstone has a mottled to clotted fabric with isolate patched clots and irregular outlines of argillaceous seams. Shale partings and pyrite stains are locally developed at the basal beds. A spotted appearance is imparted...
by recrystallized portions of sediments, showing a tan-brown diffused mottled fabric. Local interlaminated structures include tepees, flame structures and overturned microbialites. Commonly, the chaotic fabrics at the top of BI-2 grade upward into the thinly bedded microbial-laminated dolomudstone of BI-3a, with each brown lamina 0.5-1 mm thick compared to the 3-5 cm thick dark brown dolomudstone matrix. In area 1, BI-2 is not present at the basal Bass Islands Formation. The microbial-laminated dolomudstone (BI-3a) directly overlying the basal dolomitic shale (BI-1). Where BI-2 overlies the “False G Unit” in the middle Bass Islands Formation, the contact is marked by a lithologic change from anhydritic dolomudstone (BI-10) hosting desiccation cracks and crack infills to its basal medium brown, chaotic thrombolitic dolomudstone with peaked argillaceous stylolites. This contact seems gradational, with the upper BI-10 accumulates thicker dolomitic microbial laminae and thinning/disappearance of anhydrite nodules.

The brownish grey, massive bedded, microcrystalline bioturbated dolomudstone of BI-5 is restricted to Area 2 and 3. The diffused irregular brown-grey mottles typify this lithofacies (Fig 2.10-1 to 2.10-3). The mottles are throughout BI-5, creating a clotted fabric and distorting the massive dolomudstone in up-section. Portions of the brecciated zones cap this lithofacies, showing partial dissolution and re-precipitation of dolostone matrix. Megafauna are very rare, but fragments of bivalves are present (Fig 2.10-3).

The creamy, massive bedded, homogeneous dolomudstone of BI-6 are commonly pelloital, containing mm-sized anhydrite flasers and crackle fractures that have been infilled by secondary dark brownish grey dolomites (Fig 2.10-5 to 2.10-8). This lithofacies occurs regionally in the middle Bass Islands Formation that can be correlated to the basal brecciated dolostones of Raisin River Member in northern Ohio (Sparling 1970).

Lithofacies of BI-8 is restricted to area 1, rare in area 2 and absent in area 3. This lithofacies
is present in the middle Bass Islands Formation. It resembles BI-6 in its buff to light tan color and lack of megafossils. However, peaked stylolites with low relief (less than 2 mm) are very common in BI-6. Dissolution-enhanced porosity and re-precipitated dolomite are common, giving it a sucrosic structure (Fig 2.11-6 to 2.11-8). This lithofacies is commonly capped by brecciated microbial laminites and forms metre-scale shoaling cycles in area 1.

*Interpretation:* The lateral extent of BI-2 in the lower Bass Islands Formation overlying the dolomitic shale of BI-1 across the Appalachian Basin and grades into the microbial-laminated dolomudstone of BI-3 in the Michigan Basin suggests a shallowing westward depositional setting throughout the study area. During the depositional period through Salina G to the basal Bass Islands Formation, the depositional rate was low, as indicated by deposits of dolomitic shales with vertically elongate anhydrite nodules. The occurrence of BI-2 indicates a lagoonal mudflat or shallow subtidal, possibly hypersaline environment due to the presence of dissolution-collapse/dissolution pipe structures and chaotic fabrics formed by dissolution of evaporite precursors. Subangular, poorly sorted clasts and irregular vugs are throughout BI-2 that have been identified as subsurface paleokarstic features, which commonly occur adjacent to enlarged joints, fissures and surrounding caves. Near the top of BI-2, the presence of brecciated microbial laminites and caliches suggests extensive subaerial exposure that terminated the deposition of BI-2. The fabrics refer to the minerals leach from the upper layer of soil by water percolation (Flugel 2004). Plant roots take up water through transpiration and leave behind the dissolved calcium carbonate to form a brecciated texture. Though few vascular plants are thought to have established on land by the late Silurian (Kenrick and Crane 1997), as the surface dries out, capillary pressure may have driven the groundwater to flow upwards from below to form the box-like caliche fabrics in arid
areas.

The BI-5 suggests a low energy shallow lagoon or subtidal depositional environment. Intermittent carbonaceous seams and the presence of burrows all indicate normal saline shallow water conditions. No evidence of subaerial exposure has been recognized in this lithofacies, neither has that for the evaporitic fabrics.

The BI-6 in area 3 and BI-8 in area 1 all represent a low energy, continuous subtidal deposition with fluctuations in the influx of argillaceous materials. Carbonate muds in BI-8 are considered to have formed in situ and have been reworked by gastropods due to the presence of bluish-grey, diffused mottles and the low organic contents. The mottles are irregular and have been recognized as pervasive bioturbation marks. BI-6 in the Michigan Basin represents minor fluctuations of dispersal of terrigenous mud into the subtidal environments.

This facies group forms the subtidal base of metres-thick shallowing-upward cycles in the Bass Islands Formation, which is commonly overlain by peritidal deposits or evaporite-bearing successions indicative of increased arid conditions. Later reduction of terrigenous influx and return to shallow, restricted water conditions allowed for the development of microbial laminites with evaporites to form.

2.5.2 Lower Intertidal Facies (Facies Group 2)

Facies Group 2 contains lithofacies BI-7, a bituminous, very vuggy dolowackestone with intercalated microbial laminates (Fig 2.11-1 and 2.11-2). This group occurs in the middle Bass Islands Formation and is restricted to area 3. Its contact with the underlying homogeneous dolomudstone and the overlying thinly laminated microbialites are gradational, representing a complete peritidal shallowing-up cycle. Fenestral vugs are oval or semispherical in form, and are usually less than 2 mm in diameter. Vugs are partly plugged by secondary anhydrite after gypsum.
The high organic content is reflected by the bituminous dolostone matrix and the locally present concentrated carbonaceous seams. The dolomudstone matrix underlying the crinkly laminated microbialites are highly bioturbated with common dark grey bioturbation marks. No evidence of subaerial exposure was found in Facies Group 2.

*Interpretation:* Fenestral vugs in Facies Group 2 are interpreted as degassing voids within microbial communities. Gypsum has infilled the vugs selectively during the early diagenetic stage and converted to anhydrite when buried deeper. Evaporite growth has distorted the original beddings and formed chaotic fabrics in the dolostone matrix. The higher organic content indicates a reduced sedimentation rates and a relatively low-energy and restricted environment of lower intertidal zone, where the microbialite sheets were laterally developed with common fenestral vugs. The lack of macrofossils also suggests a restricted environment that was inhospitable for the development of microbial mats grazers (gastropods). This Facies Group usually forms the transitional stage from subtidal to supratidal/sabkha of cyclic, shallowing-upward successions.

### 2.5.3 Upper Intertidal to Supratidal Facies (Facies Group 3)

Facies Group 3 contains lithofacies BI-3 with BI-7 intercalations. The BI-3a and BI-3b are differentiated by the thickness of laminated beds. BI-3a consists of thinly and crinkly laminated microbial dolomudstone, and each lamina is 2–5 mm thick (Fig 2.8-1 to 2.8-5). BI-3b is thinly bedded dolomudstone, with common matrix-supported dolomite floating pebbles, and each bed is 1–3 cm thick (Fig 2.8-6 to 2.8-9). Both BI-3a and BI-3b are barren of megafossils, though locally shell fragments are found within BI-3b thin beds as allochems. The microbial laminations commonly show planar to convoluted stromatolitic fabrics (Fig 2.8-2), and are locally distorted by dissolution collapse structures (Fig 2.8-3). Tepees, desiccation cracks and caliche fabrics are very
common in up-section of BI-3a (Fig 2.8-8). Bluish grey mottles that are elongate and parallel to microbial beddings are locally developed (Fig 2.8-3). BI-3a is present at various intervals in Bass Islands Formation, capping the intertidal facies of BI-7 or the subtidal Facies Group 1. Intercalations of homogeneous dolomudstones and clayey dolostones are ubiquitous in area 3, while in area 1 and 2 anhydrite nodules and flakes are commonly present as interbeds. BI-3b is restricted to the upper Bass Islands Formation in area 3, with very common scouring structure and flat arrayed, oval dolomite pebbles (less than 2 cm in size) (Fig 2.8-7).

*Interpretation:* This Facies Group is predominant in Area 1, 2 and 3 that forms the upper shallowing-up successions of a peritidal cycle. The predominance of crinkly laminated microbialites and thinly bedded dolomudstone indicates an arid, restricted upper intertidal to supratidal or mudflat environment. The periodic exposure and high evaporation rate has resulted in the dissolution-highlighted fabrics, such as the caliche and the dewatering-induced bluish mottles, tepees and desiccation cracks. The extensive development of microbialites and absence of gastropod grazers are thought to be transitional between pre-evaporitic marine and hypersaline evaporitic conditions. Modern analogues of Shark Bay (Garrett 1970) and Persian Gulf (Kendall 1969) have the similar well developed algal mats with common desiccation cracks. Based on this analogue, it is likely that the Facies Group 3 was formed in very shallow water (1–2 m deep) with periodic subaerial exposure.

2.5.4 Restricted Lagoon Facies (Facies Group 4)

Facies Group 4 contains lithofacies BI-11. This Facies Group is the most fossiliferous in the upper Bass Islands Formation that is restricted to area 2 and 3. The medium brown, thin-medium bedded dolomudstone is highly bioturbated, with common gastropods, bivalve and brachiopod
fragments (Fig 2.11-3). Ostracods are locally abundant. Moldic porosity formed by dissolution of gastropods and ostracods is locally common. Bituminous spots are very common that contain a higher organic content against the dolostone matrix. Dark grey or brown pyrite stained stylo-seams are common, and black peaked stylolites of low relief (less than 5 mm) are rare. Facies Group 4 in area 2 is represented by abundant ostracod molds and less bituminous dolostone matrix. Often moldic porosity of ostracods coalesce along bedding planes, 2–5 mm in diameter. This lithofacies group overlies either the evaporite interbeds in area 2 or the microbial laminites of BI-3 in area 3. The contact is easily recognized by a sharp lithologic change from supratidal or sabkha dolostone to shallow water, bituminous, fossil-rich dolowackestone.

**Interpretation:** The occurrence of gastropods, bivalves and ostracods are all characteristics of a bioturbated, restricted quiet-water lagoon environment. The fauna assemblages and the bituminous dolostone matrix indicate nutrient-rich and slightly hypersaline waters. Such substrate nature is inimical to the majority of grazing fauna that feed on the organic matters within the soft sediments (Clarkson 2009). Common vugs and molds after the dissolution of gastropods and ostracods suggest the diagenetic alteration of the aragonitic shells in phreatic environment. Ostracods molds in area 2, are adjacent to the overlying dolostone unit with common enlarged joints, collapsed intervals and crackle fractures infilled by secondary dolomites. The moldic porosity may have been enhanced by the subsurface paleokarstic intervals in this way. The extent of Facies Group 4 in area 2 and 3 indicates a recurring theme seen widely in Silurian carbonates (Brunton and Brintnell 2011), and represents the last preserved deepening cycle prior to the contact of Silurian-Devonian in Ontario.
2.5.5 Shallow Stagnant Water Facies (Facies Group 5)

Facies Group 5 contains lithofacies SG-3 and BI-1. Both SG-3 and BI-1 are characterized by high shale content. BI-1 (Fig 2.6) can be differed from SG-3 (Fig 2.5-3 and 2.5-4) by containing intercalations of argillaceous dolomite clasts. The shale in SG-3 are readily distinguished from Salina F by its dark grey color and the occurrence of vertical-elongated anhydrite nodules. Shale in Salina F is greenish to pinkish grey in color and has abundant displacive light grey anhydrite nodules that are formed during diagenetic process. BI-1 contains lower shale content and the beddings are commonly distorted by dissolution of the underlying anhydrites (Fig 2.6-1 and 2.6-2). The contact of Salina G Unit and Bass Islands Formation is marked by the disappearance of anhydrite nodules and the occurrence of concentrated argillaceous seams with very common dolomite clasts floating within (Fig 2.6-1 and 2.6-4). The allochthonous dolomite fragments are oval to spherical, less than 1 mm in size, displaying various inherited characteristics form different mother rocks (Fig 2.6-1). This contact is regionally recognizable in both surface and subsurface rock records in SW Ontario, and it corresponds to a high excursion in gamma-ray logs in all cores that capture Salina G Unit and Bass Islands Formation.

*Interpretation:* Shales are commonly present in Salina Group interbedded with dolostones, anhydrites and salts (Sanford 1969). The last occurrence of shale in Silurian strata in Ontario is in the top of the Salina G Unit and the basal Bass Islands Formation. This contact seems gradational that indicates a possible cessation of subsidence and a poorly circulated water environment. Vertical-elongated anhydrite nodules have been interpreted as evaporite accumulations at the interface of sediments and sea water bottom at a low formational rate (Warren 2016), and a high salinity is thus inferred. Dolomitic shale at the basal Bass Islands Formation suggests a return to shallow water, less stagnant environment with periodic re-deposition of dolomite clasts that
overlies the Salina Group-Bass Islands contact. Recent studies have shown that the occurrence of shales is not restricted to deep-water environments and can form at virtually any water depth (Arnaboldi and Meyers 2006; Jenkyns 2010), so it is possible that the shales in Salina G and basal Bass Islands may have been formed during a hot and arid time in a shallow, sluggish ocean circulation environment.

### 2.5.6 Shallow Subaqueous Evaporite Facies (Facies Group 6)

Facies Group 6 contains lithofacies SG-3 and BI-9. Facies Group 6 is complex and contains various forms of anhydrites, which may be massive nodular, mosaic-like (Fig 2.12-3) and vertically elongated (Fig 2.5-5). This Facies Group is important for understanding the evaporite-carbonate cycles in the uppermost Salina Group and the Bass Islands Formation. Deformed anhydrite nodules are pervasive in SG-3 that caps Salina G Unit, and the vertically-elongated anhydrite nodules are present as 10-20 cm thick beds below the contact of Salina G and Bass Islands. The elongate nodular facies contain up to 60% anhydrites, leaving dolomudstone stringers or mm-sized dolomitic shales around the nodules (Fig 2.5-5). Mosaic-like anhydrites are widely distributed in “False G” unit in middle Bass Islands Formation in area 1 and 2.

**Interpretation:** The vertically elongated anhydrite nodules in SG-3 contains internal impurities and intercalations, which resembles the growth layers of the syn-depositional anhydrite nodules formed on the interface of sediments and saline water (Warren 1991). Warren and Kendall (1985) has interpreted the vertically elongate anhydrite nodules as replacement of former selenite that grew on the sediment surfaces in coastal salina environment or evaporitic lagoon. The massive and deformed anhydrites are polygenetic, and diagenetic overprints have hindered the observation of primary sedimentary features. The higher clay contents of the shale contexts that contain calcitic
or dolomitic shale interbeds, suggests a stratified and stagnant, shallow-water environment, with periodic influx of fresh water mixed with brine dilution. This is analogous to the Delaware Basin in the USA (Kendall and Harwood 1989). The occurrence of vertically elongated anhydrite nodules may represent a deepening-up cycle at the end of Salina G Unit that has experienced a restricted and poorly circulated shallow-water theme. This theme was repeated in the middle Bass Islands Formation in area 1 and 2, suggesting a more restricted depositional environment in Michigan Basin than that in Appalachian Basin while the Bass Islands Formation was deposited. More common dissolution-collapse structures in area 3, however, have been interpreted as formed by paleokarstic dissolution, which suggests a regional shallowing event during middle Bass Islands Formation was deposited.

2.5.7 Sabkha Facies (Facies Group 7)

Facies Group 7 contains lithofacies SG-1, SG-2, BI-4 and BI-10. Overall, this Facies Group has two end members: decimal- to centimetre-sized anhydrites within massive microcrystalline dolostone (SG-1 and BI-10) and the decimetre- to metre-thick beds of anhydrite nodules interfinger with microbial-laminated dolostone (SG-2 and BI-4). SG-1 is featured by centimetre-sized anhydrite nodules or crystals encased within light brown microcrystalline dolostone and microbial laminite interbeds (Fig 2.5-5). The occurrence of SG-1 marks the base of Salina G Unit that overlies the greenish grey shales in Salina F Unit regionally in Ontario. The anhydrite nodules in SG-1 have various forms, including micro- to fine crystalline, granular to spherical, lath-like to platy, irregularly arranged in the dolomite matrix. The texture of anhydrite nodules developed overgrowth fabrics after gypsum. It is uniformly 1.2–1.8 m thick in lower Salina G Unit in area 1, thins to less than 50 cm in area 2 and is absent in area 3.

Both SG-2 and BI-10 contain primary, microbial to stromatolitic features with large light grey
to bluish grey, translucent anhydrite nodules (Fig 2.12-5 and 2.12-5). The laminations are commonly consisted of 0.1–0.5 mm thick, crinkly or wavy in form. Horizons containing less than 1-cm sized laminated intraclasts are commonly found broken, with abundant tepee-like structures, desiccation cracks, degassing bluish mottles. Laminae are disrupted by large, irregular folded and deformed anhydrite nodules. The bluish anhydrite nodules displaced the dolomites, and the contact of anhydrites with the dolostone matrix is commonly diffused. The anhydrite nodules were formed by micro-aggregates or lathlike crystals. Irregular sulphate steaks are rare in the microbial laminites. BI-10 contains more argillaceous materials, and is locally present as centimetre-sized, irregular dark grey mottles in the medium grey dolostones. The mottles in the cryptocrystalline dolomitic mudstone matrix resemble the burrow trace marks. The variable size, irregular shape and the distinct contact with the matrix, however, suggests that they are not related to burrowing, but possibly to former calcium sulphate nodules decomposed by microbial activities (Dixon 1976; Kendall 1977). Massive, chicken-wire and enterolithic anhydrite are seen throughout this facies group that caps the microbial laminites. Small, uniform centimetre-sized anhydrite crystals, commonly arrayed in a row as interbeds in microbial laminites, are also present in BI-10, which is interpreted to have an eolian origin (see Warren 1991).

*Interpretation*: This facies group represents the shallowest, sabkha depositional environment in Salina G Unit and the “False G” unit in the Bass Islands Formation. It is important for recognizing the top of each shallowing-up cycle. The dolostones with nodular sulphates are well documented from modern sabkha environments, such as the Persian Gulf (Kendall and Harwood 1989). The granular and spherical anhydrite nodules in dolomite matrix, show strongly the replacement of dolomite matrix. The overgrowth structures of anhydrite nodules with microbial laminate streaks
indicate that the gypsums were formed prior to the anhydrite minerals. The inversion of gypsum to anhydrite possibly took place while the gypsum minerals were buried deep (Warren 2016). It should be noted that not all anhydrite nodules are indicative of sabkha facies (Warren and Kendall 1985). Deep burial diagenesis or replacement by evaporite minerals in ground brines can also lead to the formation of large anhydrite nodules within dolomite matrix. Typically, the sabkha facies in Salina G Unit and Bass Islands Formation interfingers with the intertidal to supratidal sediments, especially the crinkly laminated microbialites. An upper sabkha facies is represented by the occurrence of enterolithic anhydrites, which indicates a cessation of deposition and terminates one shallowing-up cycle formed in the restricted, hypersaline environment. The chicken-wire and enterolithic evaporitic fabrics overlying the microbial laminates are interpreted as formed above the water table of a fully exposed sabkha during syn-depositional or early diagenetic stage. Inversions of anhydrites and gypsum may have taken place several times prior to the lithification of those anhydrite beds (Gündogan et al. 2005). The brecciated laminites and dolomudstones, are thought to have formed by the growth and dissolution of these anhydrite nodules that have destroyed the original beddings. This facies group is closely associated with all peritidal and coastal salina facies groups at the base, which indicate progressively shallower cycles that can be regionally correlated.
CHAPTER 3. Stratigraphy and Depositional History of the Silurian Bass Islands Formation and the Bertie Formation, Southwestern Ontario

3.1 Correlation of the Bass Islands Formation and the Bertie Formation in outcrops

Besides the lower Put-in-Bay Member and the upper Raisin River Member in their type sections (Williams 1919; Sparling 1970; Liberty and Bolton 1971), the basal St. Ignace Member has been reported by Alling and Briggs (1961) in central Michigan as the basal unit of the Bass Islands Formation. In Ontario, the only outcrop of the Bass Islands Formation described by Liberty and Bolton (1971) along the Saugeen River is now covered by bank-collapse and thus inaccessible, making it difficult to correlate into the outcrop sections in Michigan.

Bertie Formation is well exposed in quarries along the Onondaga Escarpment in Niagara area (see Table 1.2 and Chapter 1.6.2 for detail). All members of the Bertie Formation are well exposed in the Ridgemount Quarry North in Bertie Township (Derry Michener Booth et al. 1989c), which has been selected as its type section in Ontario (Winder 1961). Bertie sections are also present in other quarries including Hagarsville Quarry (Standard Aggregates Inc.), Cayuga Quarry (Cayuga Materials and Construction Co.), Port Colborne Quarry (Hard Rock Paving Co. Ltd.), Nelson Quarry (Nelson Aggregate Co.) and Dunville Quarry (Dunville Rock Products Ltd.) (Haynes and Parkins 1991; Armstrong 2017; see Table 1.2).

In Ridgemount Quarry North, 11.35 m of the Bertie Formation is exposed overlain by 2.9 m thick of the Bois Blanc Formation. A 10–25 cm thick, irregular or undulatory, earthy weathered crust marks the contact of Silurian-Devonian, with very common glauconitic and phosphatic minerals (Fig 3.1). The basal contact with the underlying Salina unit is not exposed (Haynes and Parkins 1991). Greenish dark grey, fissile shale consists the upper 2 m of the Oatka Member that
is present above the water-covered pit (Fig 3.1). Selenite discs have been reported in the Oatka Member (Sanford 1969; Haynes 1989). This member is laterally continuous and can be found in all the other quarries, indicating a pervasive mesosaline environment during the initial deposition of the Bertie Formation. The dark brown, petroliferous microbial-laminated dolomudstone in the Falkirk Member overlies the Oatka Member by a sharp lithological change. Sedimentary structures including organic partings, fenestral vugs, wispy stylo-seams, degassing or dewatering dark grey mottles and desiccation cracks are well present. Algal laminitie mounds, appear in the middle of Falkirk Member, about 0.9-1.2 m high and more than 4 m across. The base of the mound overlies a 5-cm thick layer of undissolved clay residues. The top of the Falkirk Member is rich in soft-sediment deformation, including desiccation cracks, tepee structures and bleached intervals. Channel deposits of polygonal-patterns of mud chips and hopper salts have been observed. Haynes and Parkins (1992) have discovered Eurypiteris lacustris in the channel deposits in the upper Falkirk Member, which indicates the tidal channels may have cut into the intertidal to supratidal zone during the deposition of the overlying Scajaquada or Williamsville members. Scajaquada Member is uniformly 1.4 m thick and can be traced into all the quarries. It is featured by thinly-bedded grey shales and argillaceous dolostones. The contact with the underlying Falkirk is sharp and the upper contact with Williamsville Member is gradational. The absence of fossils and increased shale contents indicate a quiet and possible stagnant depositional environment. Williamsville Member is 2.3 m thick, thinly bedded, grey to brown dolomudstone and algal boundstone with shale partings. It resembles the Falkirk Member in lithology but with less desiccation marks. In western New York, this member is known for its exceptional preservation of eurypterids (Ciurca 1974; 1978; 1982; 1994; 2005; 2011). In Ridgemount Quarry North, only fragments of eurypterid telsons and carapaces are found at the basal Williamsville Member. Akron
Member is present as the topmost member in the Bertie Formation, which overlies the Williamsville Member by an undulatory and sharp contact. A thin shale layer is present 30 cm below this contact in this quarry (Fig 3.1). The Akron Member consists of wavy bedded, tan to brown mottled dolomudstone, and it reaches 3.5 m thick in Ridgemount Quarry North. It has pronounced wavy-beddings confined by irregular carbonaceous seams that are easily weathered. Irregular cavities and vugs are found partly plugged by secondary dolomites or gypsoms. The buff-mottles in the grey dolostone matrix are present in either discontinuous parallel-bedding or vertical columnar forms. Desiccation features at the top of Akron Member right below the Silurian-Devonian contact resemble the intertidal-supratidal and sabkha cycles in Bass Islands Formation. The Devonian carbonates of the Bois Blanc Formation overlies the Akron Member by an irregular, 3-10 cm thick paleokarstic surface featured by abundant undissolved clay minerals and glauconites (Fig 3.1). Glauconites have also accumulated in the solution-widened joints below the S-D contact, indicating an extensive paleokarstic interval during late Silurian to Early Devonian.

Members in Bertie Formation are traceable into other quarries along Onondaga Escarpment, though subunits within each member are not correlative. Lithofacies become more restricted and hypersaline as tracing westward into Dunville, Cayuga, Nelson and Hagarsville quarries. In Dunville quarry, 3.35 m of the Bertie Formation is exposed (Fig 3.2). The lower 1.10 m consists of the upper Williamsville Member of dark grey, fine crystalline, thin bedded shaly dolostone, with common pyrite and calcite mineralization in the vuggy porosity. It is overlain by the 2.25 m thick, medium to massive bedded, brown-grey mottled dolostone of the Akron Member by a sharp, undulatory contact. Sphalerite minerals, bituminous residues are present as partings above the lower contact. The Akron Member in turn underlies a 5.1 m thick, grey to cream, thinly microbial-laminated dolostone unit. This unit has long been formally referred to Akron Member by many
authors (e.g. Hewitt 1971). However, recent studies show that it is distinctive from the Akron Member in lithology and should be termed as a separate formation (e.g. Ciurca 1982; Haynes 1992). Ciurca (1982) named this fine-grained dolostone unit after “Clanbrassil Formation”. Based on the occurrence of the genus *Erieopterus* in this unit, he suggested it is late Silurian-Early Devonian in age, thus a younger unit than Bertie Formation. He also suggested this *Erieopterus*-bearing unit could be correlated to the Chrysler and Olney formations in central New York and the upper portion of Bass Islands Formation on the Bass Islands of Lake Erie. Based on the occurrence of microbial laminations interbedded with small algal mounds, and the very common desiccation features, Haynes (1992) proposed that this unit is the lower Bass Islands Formation. This tan to cream, microbial-laminated dolostone unit is 5.9 m thick at Cayuga quarry and 6.8 m thick at Hagarsville Quarry (Fig 3.3), all with a sharp and undulatory contact with the underlying Akron Member of the Bertie Formation. However, he interpreted this unit together with the underlying Bertie Formation as an Ohio basin facies that transgressed Onondaga Escarpment rather than a Michigan basin facies, and mistakenly correlated this succession to the whole Put-in-Bay and lower Rasin River members in the Michigan Basin.

The name “Bass Islands Formation” is favoured to refer to this above-Akron Member unit, and we suggest that it is represented by the lithofacies of BI-2 and BI-3a in the middle Bass Islands Formation above the “False G” unit as seen in the subsurface core records. Evidence comes from the lithologic similarity between this unit and the thrombolitic/microbial dolostone of BI-2 and BI-3a and the occurrence of the very common subaerial features, which may indicate a regionally extensive subaerial surface in SW Ontario following the evaporite deposition in Bass Islands Formation.
**Figure 3.1** Relationships of Oatka, Falkirk, Scajaquada, Williamsville and Akron members in the Bertie Formation at Ridgemount Quarry, Welland, Ontario.

A. Photograph showing the contact of Oatka, Falkirk and Scajaquada members in the Bertie Formation. The base of Oatka Member is covered by water, and the top of the Scajaquada Member has been eroded. Contact between each member is sharp and flat. Sedimentary breccias are restricted in the Falkirk Member.  
*(Ridgemount Quarry North)*

B. The upper two members (Williamsville and Akron) of the Bertie Formation that are overlain by the cherty dolostone of the Devonian Bois Blanc Formation. The S-D contact is marked by the lower undulatory surface above the 2-5 cm thick shale surface. The basal 30 cm of the Bois Blanc Formation is very glauconitic. Joints and coarse-grained breccias are common below the S-D contact, and some of the joints are infilled by glauconitic dolostone. *(Ridgemount Quarry South)*
A

Bois Blanc Formation

Akron Member (Bertie Formation)

Williamsville Member (Bertie Formation)

B

Scaguanada Member (Bertie Formation)

Falkirk Member (Bertie Formation)

Oatka Member (Bertie Formation)
Figure 3.2 Bertie and Bois Blanc formations in Dunville Quarry, Haldimand, Ontario.

A. Photograph showing the contact of the Akron Member (Bertie Formation), Bass Islands Formation and the Bois Blanc Formation including its basal Springvale Member. The Akron Member is readily recognizable by its mottled nature. The tan to light brown microbial-laminated dolomudstone typify the Bass Islands Formation, which overlies the Akron Member by a 2-cm thick shale layer. The contact is sharp and flat. The top surface of the Bass Islands Formation is irregular, with very common solution-collapse breccias and solution-widened joints. The Devonian Springvale Member (lower Bois Blanc Formation) is dark grey, very argillaceous sandstone with irregular chert nodules, which is in turn overlain by the light grey, very cherty dolostone of the Bois Blanc Formation.

B. Solution-collapse breccias below the S-D contact. Matrix consists of dark grey, clay-rich dolomuds. The overlying unit is iron-stained.

C. A solution-widened joint is present right below the S-D contact. Note the abundant chert nodules in the argillaceous sandstone of Springvale Member.
Figure 3.3 Bertie and Bois Blanc formations in Hagarsville Quarry, Haldimand, Ontario.

A. Photograph showing the Akron Member (Bertie Formation), Bass Islands Formation and the Bois Blanc Formation including its basal Springvale Member that have the same sequence and contact relationships with the section in the Dunville Quarry.

B. The S-D contact is well exposed on the southern wall. Note the very glauconitic nature of the Springvale Member and a pointed fold to the upper left.
Subsequently, the Akron Member is correlative to the “False G” unit in area 1 and 2 (Laurentia craton interior) and the brecciated dolostone lithofacies in area 3 (Appalachian basin). The discontinuous and wavy bedded mottles in Akron Member were previously considered as formed by bioturbation (Kobluk et al. 1977). Thin section examination shows, however, the brown mottles and the grey dolostone matrix cannot be distinguished by grain textures. The only difference is a lower content of fine grained sulphides and organic matters in the mottles. It is unlikely to be bioturbation marks and possible to be formed by the diagenetic coalescence and dolomitization of the in situ layers of dolomite nodules trapped by algal mats. Similar wavy mottles are also found in the “False G” unit (BI-10) in area 1 and 2 that seems to be formed by the destruction of previous formed calcium sulphate nodules in the algal mounds by microorganism activities (Dixon 1976; Kendall 1977). It is speculated that the Akron Member was formed in a restricted, quiet and mesosaline environment, e.g. coastal salina and the “False G” unit is formed in a similarly mesosaline but shallower sabkha environment, and these two units are thus correlative.

The subtidal-intertidal facies of the Williamsville Member shows similar shallowing-up pattern to the basal succession in the Bass Islands Formation. The dark grey shale of Scajaquada Member represents the regionally correlative shale layer that defines the basal Bass Islands Formation. The underlying thrombolitic and microbial dolomudstone of Falkirk Member could thus be correlated to the lithofacies SG-1 and SG-2 in Salina G Unit, and the greenish shale and selenite discs in Oatka Member has the same color and mineral composition of Salina F Unit. In general, only the top of the Bertie Formation is thought to be an equivalent to the Bass Islands Formation in Ontario, and is correlative to upper Salina F Unit, Salina G Unit and the basal Bass
Islands Formation. Therefore, a revised stratigraphic chart of the SW Ontario has been proposed in Figure 2.1.

3.2 Depositional Cycles in the Bass Islands Formation

The Bass Islands Formation in SW Ontario consists of metres to tens of metres thick shallowing-upward cycles. Shallowing-upward successions comprise a shallow-water, subtidal or restricted lagoon base, which is overlain by intertidal vuggy dolomudstone or microbial laminites, and capped by brining-upwards successions that indicate increased evaporation. Both carbonate and carbonate-evaporite successions have been observed throughout the Bass Islands Formation. These successions are slightly different in each area (Fig. 3.4), depending on their distinct depositional settings. In area 1 and 2, Bass Islands Formation contains more argillaceous materials, and the occurrence of “False G” unit in lower Bass Islands Formation suggested a continuous coastal-salina to sabkha depositional environment from the underlying Salina Group. In area 3, evaporites are very rare, and are present as thin interbeds with microbial laminites or replaced by secondary dolomites (e.g. dark grey, irregular mottles in “False G” unit or the tan-brownish grey mottles in Akron Member of the Bertie Formation). This lateral lithofacies change from open marine to restricted lagoon in area 3 to more restricted lagoon and sabkha variations enables a reconstruction of an inclined ramp in SW Ontario during late Silurian. These brining-upward successions can be recognized as shallowing-upwards cycles in the Bass Islands Formation. These genetically related cycles are relatively conformable and are bounded by lower marine-flooding surfaces and upper subaerial or karstic surfaces. Salina G Unit consists of one or two brining-upward cycle regionally, and multiple metre-scale to a few metres thick cycles are readily recognizable in the Bass Islands Formation.
Typical thickness of Salina G is 3-8 m. In area 3, it is uniformly 3-3.5 m thick, and thickens upto 8 m in area 1 and 2 southwestward. The lower cycle of the Salina G consists of intertidal-supratidal carbonates with common replacive evaporate laths overlain by mosaic-like anhydrites or thick bedded, bluish grey anhydrite nodules bounded by microbial laminites. The upper cycle comprises of basal dolomitic shale with replacive anhydrite nodules overlain by subaqueous, vertically-grown anhydrites (Fig 3.4E).

Four types of carbonate shallowing-upward successions have been recognized (Fig 3.4 A-D) in the Bass Islands Formation and two of them are capped by karstic tops (Fig 3.4 C and D). Regionally, the base of the Bass Islands Formation consists of 3–5 m thick Succession A. The basal shale or dolomitic shale is overlain by thrombolitic dolomudstone and laminated microbial dolomudstone (BI-2 and BI-3). Intraformational clasts are commonly found capping the cycle. Succession A is present completely in area 3 with a uniform thickness of 3-5 m. The basal dolomitic shale thins southwestward into area 1 and 2, and pinches out in area 1 in Bruce and Huron counties. In area 1, Succession A is incompletely present as tens of centimetres thick basal shale overlain by 3-7 m thick thrombolitic dolostone and laminated microbial dolomudstone.

In area 1 and 2, cycle A is overlain by the “False G” Unit that is represented by Succession F and G (Fig 3.4). Succession F is indicative of a complete carbonate-evaporite parasequence formed in peritidal-sabkha transitional environment. The basal Succession F consists of various subtidal lithofacies types (Facies Group 1). It shallows upwards into the intertidal, bituminous dolomudstone with fenestral vugs that are partly plugged by anhydrites overlain thrombolitic and microbial dolostone. Succession F has microcrystalline dolomite with mosaic-like or chicken-wire anhydrites at its top. The overall Succession F suggests an increased restriction and higher evaporation shallowing parasequence. Succession G is consisted of the lower dolomudstone with
irregular dark grey mottles and the upper subaqueous vertically elongated anhydrite nodules, which are sometimes overlain by mosaic-like anhydrites and anhydrite nodules encrusted by microbialites. In area 3, “False G” unit is absent and a paleokarstic interval overlying the Succession A (see Chap 2.9) is interpreted as equivalent to “False G” unit in area 1 and 2.

In area 2 and 3, the middle Bass Islands Formation consists of several complete or incomplete cycles of Succession B and C. In Succession B, subtidal bioturbated or homogeneous dolomudstone is overlain by intertidal bituminous microbial laminites. Supratidal thromboilite and laminated microbial dolomudstone commonly caps this cycle. Present in upper Bass Islands Formation in area 2 and 3, Succession C is similar to Succession B, with a distinct gastropod-sucrosic dolostone base and a common karstic surface featured by dissolution joints and breccias under the Silurian-Devonian contact.

In area 1, overlying the “False G Unit” the middle-upper Bass Islands Formation consists of several (6 to 8) carbonate cycles of succession D that may or may not have a karstic top. Succession D is similar to Succession B, but with a more argillaceous base and much thicker intertidal-supratidal microbial-laminated dolomudstone that occurs in the middle-upper Bass Islands Formation.

Because of the very shallow water depositional environment and the pervasive karstic events during the deposition of Bass Islands Formation, a slight change of sea level could thus severely change the lithofacies distribution. Therefore, subunits within each succession (Fig 3.4) are difficult to correlate among subsurface cores.
Figure 3.4 Generalized shallowing-upward cycles in Bass Islands Formation and Salina G Unit. Both carbonate and carbonate-evaporite successions are recognizable. See Fig 1.7 for legends.
3.3 Depositional History and Regional Correlation of the Bass Islands Formation

The depositional history of Salina G Unit and Bass Islands Formation is reconstructed on the stratigraphic architecture and shallowing-upward patterns of lithofacies groups listed in Table 2.1, integrated with correlation of subaerial exposure surfaces and paleokarstic intervals. Cross-section A-A’ and B-B’ best illustrate the vertical stacking patterns in the Bass Islands Formation on the western Appalachian Basin and eastern Michigan Basin in SW Ontario (Fig 3.5 and 3.6). The shale at the top of Salina G Unit is chosen as a datum of correlative bed because it is a flat-lying, regionally extensive unit that has a distinct log signature. In area 3, the Bass Islands Formation represents a restricted mudflat-peritidal depositional environment. Evaporites are rare in area 3, though karstic features formed by dissolution of pre-existing evaporites have been recognized.

Lithofacies distribution indicate that the Bass Islands Formation was deposited in a continuous ramp during late Silurian, and the thickness increases westward towards the craton interior. The lack of wave-related fabrics indicate that the study area was possibly located in a protected carbonate ramp. This thickness variation is likely to be controlled by the flexure of forebulge region during the late Silurian (Ettensohn and Brett 1998) that has uplifted the Niagara Peninsula area and the adjacent western New York. This regional uplift has eroded the upper Bass Islands Formation overlying the mottled dolostone in Akron Member of the Bertie Formation in Niagara area and the adjacent New York state.

Based on the stacking patterns of the different lithofacies (Fig 3.4), the mixed carbonate-evaporite successions of the Bass Islands Formation can be grouped into three depositional cycles. The transgressive stage is represented by relatively deeper facies (restricted lagoon/subtidal-lower intertidal). These facies are commonly heterogeneous and are composed of various types of lithofacies (e.g. BI-2, BI-5, BI-6, BI-7 and BI-8). The highstand systems tract is represented by
the infills of the vertical accommodation space by microbial-laminated dolomudstone. The regressive stage is marked by supratidal deposits and sabkha evaporites. The extensive subaerial exposure surfaces and the micropaleokarstic features below usually marks the boundaries between each cycle.

3.3.1 Cycle 1 (lower to middle Bass Islands Formation)

The lower Bass Islands Formation consists of two shallowing-up successions (A and B in Fig 3.4). The basal succession A consists of 3-5 m thick, regionally traceable, peritidal shallowing-up succession (Fig 3.4A). The shale unit in the underlying Salina G Unit represents the maximum flooding surface, and it characterizes a regional restricted, possibly shallow stagnant and anoxic environment across from Appalachian Basin into Michigan Basin. The faunally barren dolomitic shale at the base of Bass Islands Formation and the equivalent Scajaquada Member in Bertie Formation was deposited at the initial shallowing-up stage under emergent conditions in a shallow, poorly circulated water realm. In area 2 and 3, this facies grades upward into the subtidal thrombolitic dolomudstone with very common dissolution breccias and geopetal features that are partly infilled by evaporites. It is in turn overlain by the intertidal-supratidal microbial-laminated dolomudstone with a brecciated microbialite lithofacies with very common desiccation cracks and caliche textures. In area 1, the dolomitic shale at the basal Bass Islands Formation is overlain by the microbial-laminated dolostone directly, and the succession is capped by the brecciated/recrystalline dolostone with common solution-collapse breccia.

In area 3, the upper succession is a 2-8 m thick, peritidal-sabkha interval overlying the Succession A regionally. The basal boundary is marked by an abrupt lithological change from the subaerial lithofacies top of Succession A to the overlying subtidal facies group that represents a transgression following the exposure interval. The transgressive subtidal dolomudstone is either
overlain by 1-3.5 m thick microbial-laminated dolomudstone, which is capped by a brecciated unit; or by the equivalent mottled dolomudstone in the Akron Member of the Bertie Formation. Paleokarstic cave systems have been recognized below the upper boundary, as cave floor clasts, solution-collapse breccias, cave deposits (speleothems) and crackle breccias at the cave floor (see Chapter 4 for detail).

In area 1 and 2, the microbial-laminated dolomudstone is overlain by thick successions of carbonate-evaporites (“False G” unit) that represents the shallowest water deposits. It represents the shallowest sabkha-salina deposits as the result of regression in the middle Bass Islands Formation. The top cycle boundary is marked by the disappearance of anhydrite nodules, which is correlative to the boundary of the Put-in-Bay Member and the Raisin River Member in Michigan and that of the top of the Akron Formation in New York. In the outcrop belt along western New York and the adjacent Niagara area, the Akron Member of the Bertie Formation is overlain by the Devonian Oriskany Formation or the Bois Blanc Formation, therefore the overlying Cycle 2 is absent because of the presence of erosion or non-depositional hiatus during late Silurian to Early Devonian (Boucot 1968).

3.3.2 Cycle 2 (middle to upper Bass Islands Formation)

Cycle 2 is defined by a carbonate-dominated succession consisting of multiple superimposed lower-level parasequences (Fig 3.5 and 3.6).

In area 1 and 2, the boundary between Cycle 1 and 2 is marked by the disappearance of anhydrite nodules at the top of “False G” unit. In area 3, this boundary is marked by a regionally extensive cave system with common solution-collapse breccias and crackle fractures. The lower transgressive part of the cycle consists of several shallowing-up peritidal successions (C and D in Fig 3.4). Each succession is exclusively capped by a brecciated interval with very common
solution-induced breccias and fenestral vugs. Units within each succession are not regionally
correlative because of the very shallow depositional environment and the irregular
paleotopography during the deposition of Bass Islands Formation. In area 1, the relative deepest
water deposits comprise the creamy homogeneous dolomudstone (BI-6). In area 2 and 3, the
relative deepest water deposit is presented as the bituminous dolowackestone with sparsely
scattered bivalve or gastropod fragments, with occasional dolomite flat pebbles and very common
carbonaceous seams. These deposits gradually transition upward into the subtidal dolomudstone
and the peritidal microbial-laminated dolomudstone under the Silurian-Devonian boundary, under
which solution-widened joints, solution-enhanced vugs and cavities that are partly plugged by
secondary dolomite, collapse breccias and joints infilled by Devonian quartz arenite are very
common.

3.3.3 Cycle 3 (uppermost Bass Islands Formation)

Cycle 3 is only present in area 1 and 2 on the Laurentia craton interior. In area 3, Cycle 3 is
not present, indicating a differential erosion of during the S-D contact. The topmost of the Bass
Islands Formation usually consists of cream to buff, homogeneous dolomudstone with rare
gastropods. It marks a return to shallow water depositional environment, though top of the Bass
Islands Formation has been eroded. Therefore, this cycle is not completely preserved in Ontario.
In New York, the Rondout Formation overlying the Bertie Formation is featured as massive, vuggy
and heavily bioturbated dolostone and brachiopod-coral dolowackestone (Brett et al. 2000). These
two units all represent a regional deepening of sea-level followed the deposition of the peritidal-
sabkha evaporite-carbonate successions prior to the regional uplift. Whether the upper Bass Islands
Formation can be correlated to the Rondout Formation is unknown. However, an interval of
ostracod-brachiopod dolostone unit is present overlying the top of the Bass Islands Formation in
area 2 (Fig 3.5). A good example comes from the Core 19 (Consumers’ Amoco 13076). This core locates 24 km north of the Point Pelee in area 2. The dolostone unit is 14 m thick, consisting of common ostracods, brachiopods, bivalves and gastropods. On top of this unit, a 10-cm thick, dark grey, stromatoporoid-coral bafflestone is present right below the prominent unconformity. This lithofacies has not been reported within the Bass Islands Formation in previous studies. Brett et al. (2000) has described the same fauna in Rondout Formation in New York. The Rondout Formation is a massive, bioturbated dolostone unit with brachiopod and coral fossils. Stromatoporoids have been reported in the Manilius Formation (Ciurca 1990). Therefore, the Silurian-Devonian conformity is thus speculated to be present in the subsurface cores in Area 2, which is overlain by the Early Devonian Wallbridge unconformity (Sloss 1963; Dennison and Head 1975). Further studies are required to study the stratigraphic relationships of the Bass Islands Formation and the Rondout and Manilius Formations in New York.

3.4 Summary

The lower Bass Islands Formation SW Ontario consists of cyclic carbonate-evaporite successions and the upper Bass Islands Formation consists several lower-level lagoon-peritidal carbonate successions (Fig 3.6). In Niagara Peninsula, the Scajaquada-Willimsville-Akron members in Bertie Formation are correlative to the Cycle 1 of Bass Islands Formation. The peritidal-sabkha facies dominate the lower Bass Islands Formation indicating a widespread deposition in a shallow evaporitic ramp in Area 1 and 2 and inner ramp in Area 3. The lagoon-peritidal facies in Area 2 and 3 and the peritidal dominated lithofacies in Area 1 suggest an easterly inclined inner ramp depositional environment. Karstic features are throughout the Bass Islands
Formation that usually cap each high-level cycle, indicating a significant time break between the exposure and the subsequent transgressive stage.

The Bass Islands Formation reaches the thickness of more than 75 m in area 1 and 2, and is interpreted as formed along the eastern flank of the Michigan Basin with a rapid evaporite accumulation rate in the “False G” unit. It thins to less than 15 m in Area 3, which is interpreted as a lower carbonate formation rate (possible more than 10 times slower than the evaporite accumulation rate), and a differential erosion of the top of Bass Islands Formation by the movement of the forebulge region (Ver Straeten and Brett 2000). It is also possible that the gently inclined inner ramp has a low carbonate productivity in Area 3.

In the study area, paleogeographic reconstruction of the SW Ontario indicate the edge of the sabkha-salina facies likely to reflect a paleoshoreline parallel to the modern western shoreline of Lake Erie. However, the lack of cores in the Algonquin Arch region has prevented the regional correlation from Appalachian Basin to Michigan Basin. Though direct contact relationships of Bass Islands Formation in Area 1 and 3 are not accessible, indirect evidence from facies analysis indicate the southwestern elongated shoal complex is situated west of the Long Point, where Haynes (1992) discovered the Bass Islands Formation started to overlie the Bertie Formation.
Figure 3.5 Stratigraphic cross section A-A’ showing the lithofacies distribution and depositional sequence of the Bass Islands Formation below the S-D contact from Niagara through along the western Lake Erie to Essex County. (see well locations in Fig 2.4).
Figure 3.6 Stratigraphic cross section B-B’ showing the lithofacies distribution and depositional sequence of the Bass Islands Formation below the S-D unconformity from Bruce County to Lambton County (see legends in Fig 3.5 and well locations in Fig 2.4).
CHAPTER 4.  Paleokarst in the Bass Islands Formation, Southwestern Ontario

4.1 Introduction

Paleokarstic features in evaporitic-carbonate successions are very common in many sedimentary basins around the world (Friedman 1997), as well as during non-depositional hiatus when complex paleokarstic systems are commonly formed by chemical dissolution in both surface and subsurface terranes (Fritz et al. 1993). This diagenetic feature is defined as an “overprint in subaerially exposed carbonate bodies, produced by dissolution and migration of calcium carbonate in meteoric/phreatic waters and generated a recognizable landscape” (Roehl 1967). The Bass Islands Formation in SW Ontario collectively display an extensive of previous unrecognized subsurface paleokarst features (Brunton and Dodge 2008), evidenced by tufa deposits, disappearing streams, breathing wells, and spring discharge along major rivers. Significant overburden of last ice-aged glacial deposits has covered the Bass Islands Formation outcrop belts and makes it difficult to access the karstic terrain in Huron and Bruce counties in SW Ontario (Brunton and Dodge 2008; Carter and Brunton 2011). The prolonged exposure at the boundary of late Silurian into Early Devonian resulted in multiple stages of karstification by the dissolution of subsurface evaporites or carbonates in both vadose and phreatic zones.

Previous studies of the paleokarst profiles in SW Ontario are confined to the Niagara area. Surface dissolution-induced karstic features in Niagara Peninsula area, including solution-widened joints, leaching, small-scale sinkholes and boring traces of Trypanites have been documented by Kobluk et al. (1977) and Pemberton et al. (1980) below the Silurian-Devonian contact. Hayens (1992) has documented various types of breccias in Bertie Formation along its outcrop belt in the
Niagara area that corresponded to syn-depositional exposure or later Acadian tectonic cycle in the Middle Devonian. In the USA, Carlson (1992) has reported the paleokarstic features in late Silurian units in outcrops as a result of regional uplift during earliest Devonian in northern Ohio. Black (1997, 2012) has interpreted the paleokarstic system below the Silurian-Devonian contact in Michigan as front zones of evaporite dissolution in the peripheral area of Michigan Basin.

However, no detailed subsurface studies have been carried out on the paleokarstic features in Ontario. In the cored wells that capture the Bass Islands Formation, macroscopic dissolution-collapse structures are not well identified because of the limited scale of view. Subsurface paleokarstic features are then inferred by the recognition of systems of molds, vugs, in situ breccias, regional leaching, solution-enlarged joints, boxwork, buried caves, and cave deposits. As inferred by the lithofacies interpretation (see Chapter 2), both during and after the deposition of the Bass Islands Formation, large volumes of evaporites and carbonates have been possibly dissolved to form solution-collapse breccias, cave systems and sinkholes.

4.2 Paleokarst Features

Based on core logging and field observation, the karst features in the Bass Islands Formation are divided into four categories in this study.

The first group occurs at the top of the Cycle 1 at the lower Bass Islands Formation, comprising collapsed cave systems, such as discrete brecciated microbial or homogeneous dolostone beds overlain by cave deposits and crackle breccias in cave roof. These beds range from less than 50 cm to more than 5 m. Laterally extensive layers of brecciated algal mats are present in this group, which potentially can be confused with the channel infills or storm beds, but can be distinguished by the coarse cements that maintain the pseudomorphs of the pre-existing evaporites.
The second group occurs in Salina G Unit and the middle-upper Bass Islands Formation, in the form of syn-sedimentary and/or early diagenetic breccias. These beds are present as either channel infills or subsurface meteoric solution-collapse breccias. Desiccation cracks have also been recognized throughout the middle-upper Bass Islands Formation, in the form of intraformational clasts, tepees and mud chips. These beds are commonly found tens of centimetres below a subaerial surface and have poor stability for regional correlation.

The third group occurs right below the Silurian-Devonian unconformity, which could sometimes extend tens of metres downward into the middle Bass Islands Formation (Carlson 1992). Near the unconformity, breccias and solution-widened joints are commonly formed at the atmospheric-rock interface by meteoric water percolation and in shallow phreatic zone. Subsurface paleokarsts may have undergone groundwater drainage to form the underground caves during the regional uplift in the Early Devonian. Devonian siliciclastic sandstones from the Oriskany Formation or the Springvale Member of the Bois Blanc Formation, are discovered metres below the contact as cave or paleokarstic conduit infills that have been interpreted as having been transported by groundwater through subsurface karstic systems.

The fourth group include burial pressure-dissolution breccias surrounded by small-scaled peaked stylolites. It is usually clast-supported with higher argillaceous material contents. The irregular and poorly sorted breccias in this group are of the same lithology and are surrounded by the argillaceous stylolites that represent the undissolved residues. The study area has undergone several tectonic subsidence to uplift cycles resulting in the dissolution of the deep buried carbonates and evaporites during Middle Devonian to Carboniferous time (Boucot 1968; Johnson et al. 1992). The burial breccias may have formed either during the Middle-Late Devonian Acadian Orogeny or even during the Pleistocene glaciation when the study area was covered by recent
glacial drifts (Haynes 1991; Black 2012).

4.2.1 Cave Systems in the lower Bass Islands Formation

Cave systems are well developed in the lower Bass Islands Formation below the top of Cycle 1. Three breccia types are commonly found formed by cave collapse and possible tectonic fracturing. These solution-collapse breccias play an important role in regional stratigraphic correlation and development of groundwater reservoirs. The three types of breccias include the cave roof crackle fractured breccias, the solution-collapse cave infills and the lower cave floor rubble breccias.

A good example of this cave system comes from Core 2 (U.S Steel No.1) from 156–158.5 m (Fig 4.1). The cave roof karstic lithofacies consists of unbrecciated, homogeneous dolomudstone at the top grading downward into crackle fractured breccias (Fig 4.1C). Cave roof dolomudstone is commonly leached by meteoric water or fresh groundwater, with a high porosity ranging from 2-20%. The crackle fractured breccias are rectangular with little displacement that have been interpreted as formed at the incipient brecciation stage in an extensively permeated dolostone layer where the dolostone fragments have not been totally dissolved and collapsed. These breccia types with common centimeter-scaled fractures are usually infilled by argillaceous dolomites, meteoric cements, secondary saddle dolomite or evaporites (Fig 4.1A). These breccias resemble the tectonically fractured breccias, but have a less extensive spatial distribution and are commonly found overlying the dissolution-collapse breccias or cave deposits (e.g. speleothems). The enhanced porosity and the fissures make the cave roof facies excellent reservoir rocks for potable groundwater or hydrocarbon.

Collapse clasts as cave infills are usually formed by structural collapse of cave walls or cave roofs in a previously open cave or cavern network. The clasts are commonly irregular in form,
Figure 4.1 Karsts in Core 2. U.S Steel No.1. Karstic features are readily visible in core No.2. Upper interval directly beneath the S-D contact shows a succession of epikarst features. Cave system consists of features shown in A-C.

A. Cave floor. Irregular, subangular microbialite clasts are cemented by recrystallized dolostone. Replacement of pre-existing evaporites and laminite breccias.

B. Cave infills. Dark grey shales and a caliche unit overlie the cave floor breccias sharply. Shaly contents may be the undissolved residues in cave.

C. Cave roof. Mosaic dolostones and crackle fractures are present, formed by gravity and pore water dissolution.


E. Oriskany siliciclastic sandstone clasts. Subangular clasts of the Devonian quartz sandstone incorporated into the Silurian dolostones. A karstic conduit enables the sand grains to infiltrate downward into the Silurian strata.

F. Bleached dolostone. Massive, very porous, light buff dolostone. Porosity is highlighted by meteoric dissolution.

See Fig 1.7 for legends.
poorly graded and may have various carbonate or evaporite lithologies. Solution-collapse breccias are characterized as commonly clast-supported with interstitial cement or clayey matrix (Fritz et al. 1993). The matrix supporting the breccias may have been compromised by the dissolution of cave floor carbonates/evaporites. Open spaces among framework clasts are commonly infilled by undissolved clay residues or shales. Centimetre-sized geopetal structures are locally developed (Fig 4.1B). In the vertically elongated, irregular geopetal structure, the bottom is infilled by microbial laminated dolomudstone and the top is plugged by euhedral sparry dolomite. When the host rock is internally dissolved, gravity collapse can lead to the formation of multi-layers of breccias (Fig 4.2-9). In Core 2. U.S. Steel No. 1 (Fig 4.1), the cave infills consist of the heterotic rubble collapse breccias as thick as 10 cm. The cave system extends to 1.6 m in height. The collapse breccias exhibit dominantly random chaotic fabrics with fractured clasts, and the matrix is dark grey, clayey dolomudstone. A 7-cm thick shale layer is present right above the cave floor, which may represent the undissolved residues. Below the cave floor, the microbial-laminated dolomudstone shows a brecciated texture that may have resulted from the dissolution of evaporites in host rocks. Evidence of previous presence of evaporites also comes from the presence of a 5-cm thick anhydrite layer right below the microbial-laminated dolomudstone facies.

The surface of the cave floor is usually undulatory or irregular. The breccias consist commonly of the same lithology as the underlying host rocks, but have meteoric cements observable in thin sections. In some cores the cave floor breccias show a finer-up grading pattern. In other cases, cave floor can also serve as the cave roof of the underlying cave system (Fig 4.2-7 and 4.2-9).

In general, subsurface cave systems are commonly found in the lower Bass Islands Formation, with distinctive features of cave breccias. These cave systems may have been formed during the
exposure period between Cycle 1 and Cycle 2. Enhanced fluid flow associated with tectonic activity during the earliest Devonian may have rejuvenated the cave systems as well. The dissolution of carbonate usually starts in the vadose zone by meteoric water percolation, but can be pronounced in the phreatic zone when the brine level lowers during major tectonic uplift. Fresh water could thus enhance the dissolution and collapse structures of the previous cave systems and form an extensive karstic subsurface terrain (Fig 4.6 and 4.7).

However, the continuous uplift of the study area during the Salinic Orogeny/Disturbance (Boucot 1968) may lead to multiple levels of phreatic zones and the continuous lowering of brine levels, so caves were likely to form or enhance in various intervals of the Bass Islands Formation. These caves may or may not be interconnected, but the tabular phreatic caves infilled by undissolved residues show a regionally correlative high response to gamma ray log and high response to neutron porosity, which have a distinguished signature from the depositional shale layers. These undissolved shale residues on the cave floor can thus serve as aquitards to better confine the overlying groundwater flows, and a regional aquifer system may have developed within the cave infills (see Fig 4.7).

4.2.2 Collapse Breccias by Evaporite-Solution

Both the Salina G Unit and the “False G” unit in the middle Bass Islands Formation have large proportions of anhydrites (20–70%). The dissolution of anhydrite or gypsum may collapse and generate solution-collapse breccias of carbonate clasts (Fig 4.2-1 to 4.2-4), which plays an equally important role in the formation of subsurface karstic caverns and sinkholes (Friedman 1997). The overgrowth of the displacive evaporite nodules usually destroy the original sedimentary beddings and structures and form chaotic fabrics within the dolostone matrix (Fig 4.2-3). Angularity of the clasts is common accentuated around the anhydrite nodules. Sulphates dissolve
Figure 4.2 Paleokarstic features in the Bass Islands Formation. 1-4: evaporite-solution induced breccias. 5-9: solution-collapse breccias and sedimentary breccias.

1. Evaporite-solution breccias. Note the cycles of brecciated microbialites distorted by the overgrowth of anhydrite nodules. Anhydrite precursors were replaced by secondary dolomites. *(Core 2. U.S. Steel. No.1)*

2. Evaporite-solution breccias. Note the evaporite growth around the dolomite breccias. A possible buried sinkhole system with irregular floor and oval-clast breccias as sinkhole infills *(Core 4. Consumers' Amoco 13061)*

3. Incipient dissolution of evaporites. Small, irregular vugs are formed by the dissolution of replacive anhydrites. Some show geopetal structure consisting of lower dolomite clasts and upper dolomite cements *(Core 8. Imperial No. 809 - W.J. Mawson No. 1)*.

4. Brecciated fabrics in the lower. Note the pseudomorphs of selenite minerals that have been replaced by dolomites *(Core 11. OGS 82-3)*.

5. Sedimentary breccias below a subaerial exposure. Caliche-like fabrics are well developed by the dissolution of meteoric water dissolution process near the exposed surface *(Core 4. Consumers' Amoco 13061)*

6. A buried cave system. A 2-cm thick crackle fractured breccias near the top represent the cave roof. Solution-collapse breccias fell along the subvertical joint in the middle. Cave floor is 35° inclined. *(Core 7. BH08-21)*.

7. An overprint of two karstic intervals. The upper breccias consist of laminated microbialites, and the lower breccias are of various lithologies. *(Core 19. Consumers 33409)*.

8. A large cavity that has been infilled by secondary dolomite *(Core 22. Cansalt DDH 87-3)*.

9. Two brecciated intervals. The upper brecciated interval is features as caliche structure. A joint connects the caliche deposits downward into the cave system. Solution-collapse breccias are of various dolostone lithofacies. *(Core 9. Consumers 13076)*.
rapidly to form karstic features at a rate about 100 times faster than dissolution of limestone (Cooper 1996). Modern analogues show that a 3-m cube of gypsum dissolution can form a cave within 18 months (Warren 2006).

In the study area, evaporite-dissolution features are commonly associated with the brecciation of microbial-laminated dolomudstone (Fig 4.2-1). Breccias in the thrombolitic dolomudstone (BI-2) and on top of “False G” unit are thought to be formed by the dissolution of the underlying evaporites. Fig 4.2-3 shows the incipient replacement of evaporites within the microbial dolomudstone matrix. During early diagenesis, the unaltered microbialite beddings may act as aquitards and confine the groundwater to flow within the evaporitic zones (Fig 4.6-1). Most evaporite minerals are highly soluble in fresh water and the bacterial or thermal reduction of the evaporites is effective in accelerating the dissolution of anhydrite or gypsum (Friedman 1997). The dissolution rate is controlled by the groundwater chemistry and flow velocity. The evaporites tend to dissolve rapidly by the surrounding meteoric water and form a collapse-solution fabric (Fig 4.2-1). In the phreatic or mixing zone that is supersaturated in respect to carbonate with a normal salinity, the evaporites tend to be replaced by carbonate minerals and form chaotic fabrics showing pseudomorphs of the evaporite precursors (Fig 4.2-1 and 4.2-2). In thin sections, the dolomite cements that replaced the evaporite precursors show a double-termination or tapered feature. The dissolution, reduction and replacement of evaporites commonly results in extensive vuggy porosity and solution-collapse structures, even when there is no relict of evaporites left (e.g. in the Falkirk Member, Bertie Formation).

The evaporite-solution collapse breccias have been widely recognized (Hayens 1991), but the age of the evaporite dissolution is hard to determine. Multiple episodes of brecciation may have taken place during the deposition of the Bass Islands Formation and even during the Middle-Late
Devonian Acadian Orogeny (Smith 1992). Haynes (1991) and Carlson (1992, 1994) reported vertical evaporite-solution pipes extending 30 m through the Bass Islands Formation in the Fox Quarry in southwestern Michigan. Carlson (1992) noted the solution pipe cuts the interstratal breccias and are infilled by Sylvania glauconitic sandstone, based on which he suggested the evaporite-dissolution processes may have taken place during the deposition of the Middle Devonian Lucas Formation (Detroit River Group) or even during the tilting of the study area in the first phase of Acadian Orogeny in the Middle Devonian. It is likely that the dissolution of the subsurface evaporites have created more accommodation room for the Early-Middle Devonian carbonate depositional systems (Brunton and Brintnell 2011; Brintnell 2012).

4.2.3 Sedimentary Breccias

The sedimentary breccias usually consist of matrix-supported, rounded or angular carbonate clasts that are formed by syn-sedimentary or early diagenetic processes (Blount and Moore 1969). During the deposition of the Bass Islands Formation, periodic erosion of the exposed carbonate terrain has created complex dissolution fabrics including desiccation cracks, tepees, caliches, dewatering structures and collapse breccias in the vadose zone above water table (Fig 4.6-3). In the middle to upper Bass Islands Formation, angular carbonate clasts of homogeneous dolomudstone or microbial-laminated dolomudstone within a dark brown, clayey dolomudstone matrix are commonly observed right below caliche deposits (Fig 4.2-1 to 4.2-5). Solution joints usually form along vertical or sub-vertical zones, and the carbonate clasts along the joint edges may have fallen into the joint system (Fig 4.2-6). The angular clasts are usually of the same lithology, indicating that the joints are locally developed on a smaller scale and the breccias are formed by the collapse of the overlying unit that is no more than 1 m above. The sedimentary breccias formed by meteoric dissolution resemble storm breccias, but show a poorly graded pattern
(Fig 4.2-6 and 4.2-7). Meteoric meniscus cements have been observed in the dolomudstone matrix that supports the carbonate clasts. Noncontemporaneous breccias have been recognized by the presence of internally fractured clasts. The clasts exhibit some rough imbrication along the edges. These breccias may have been formed as bank-collapse to infill channels. Wispy argillaceous styloseams, are commonly found surrounding or intercutting the clasts that have been interpreted as deep burial dissolution features in late diagenetic process.

The lateral extent of the sedimentary breccias is hard to determine by core observation, but the vertical extension is usually within tens of metres below the S-D surface. In the outcrop, Hayens (1991) has reported a 2-m high and 50-m wide sedimentary breccias in the Cayuga Quarry. Size the sedimentary brecciated zones vary from 0.3-5 metres that caps each shallowing-upwards succession. They may have formed a complex subsurface aquifer network in the middle to upper Bass Islands Formation. These sedimentary brecciated zones play an important role in forming the portable water aquifers near the surface in Niagara area, though individual subsurface channel is hard to predict.

The diagenetic processes, including desiccation and dewatering/degassing of the microbial mats, have created various types of breccias, such as desiccation crack breccias, tepees and mud chips. These breccias are present in many microbial-laminated dolomudstone units within sabkha mudflat sequences. In the logged cores, desiccation cracks are usually centimetres long, and less than one centimetre wide. However, in outcrops (e.g. Ridgemount Quarry North, Fig 3.1), desiccation cracks up to 20 cm wide and slump-rotation of blocks are identified with broken-off dolomudstone clasts (Fig 3.2). Tepees are present associated with the desiccation cracks, and are of small lateral extent. Irregular clasts infilled the polygonal interspace and form a connected permeable zone. These brecciated features may result in poor ground stability and form an irregular
subsurface void system, adding up the hazardous potential of surface collapse along the Bass Islands/Bertie cuesta in SW Ontario.

4.2.4 Deep burial breccias

The deep burial breccias in the Bass Islands Formation are difficult to classify because most of them have undergone multiple phases of dissolution. These brecciated intervals are marked by highly irregular, clast-supported layers of porous boxwork (Fig 4.3-1 and 4.3-3). The breccias consist of irregular, subrounded microbial-laminated dolomudstone, with a vuggy and crystalline matrix where present, and are surrounded by the peaked argillaceous stylolites (Fig 4.3-1). Deep burial breccias are also well developed within the thrombolitic lithofacies (BI-2) and the Salina G Unit. Open spaces may contain microbial-laminites, secondary dolomites and anhydrites (Fig 4.3-3). These breccias are best developed in the Falkirk Member in the Niagara area, extending into the lower Salina G in the remainder of SW Ontario. This brecciated zone is confined by the underlying shale of Salina F and the overlying dolomitic shale of the basal Bass Islands Formation, and forms a regionally consistent bedrock sulphate aquifer (Fig 4.7). These breccias may have been formed by the dissolution of replacive evaporite, and enhanced by deep burial pressure-solution and reactivated during regional uplift. The ancient and modern hydraulic fractures may have facilitated the development of pathways within this interval that have dissolved large amount of evaporites and transported them into the subsurface structural topographic lowlands (Black 1997; Carter 2016).
**Figure 4.3** Pressure-dissolution breccias (1-3) and the paleokarstic features under the S-D contact (4-9).

1. Clast-supported breccias surrounded by argillaceous stylolites. Vugs are very common. (Core 1. DOMTAR GODERICH S.T.#1).
2. Clasts of dolomites are highly recrystalline. Argillaceous stylolites are horizontal. Chaotic breccias are present in the upper unit. (Core 9. Consumers' Amoco 13076).
3. Clast-supported breccias surrounded by argillaceous stylolites. Open spaces among breccias are infilled by undissolved clay residues. (Core 11. Consumers' Pan Am 13057).
4. Layered speleothems intercut by a solution widened joint 10 cm below the S-D contact. (Core 10. Consumers' Pan Am 13057).
5. A solution-collapse feature that has been infilled by dolomite breccias at the bottom and Devonian siliciclastic sands at the top. (Core 13. ARGOR 65-1).
6. Joints and vugs are highly enhanced by dissolution below the S-D contact. (Core 22. Cansalt)
7. Sinkhole infills of quartz arenite matrix and dolomite clasts above the S-D contact. Dolostone clasts are of various lithofacies, that have been interpreted as collapse breccias from surrounding walls. At least four collapse events are recognizable. (Core 7. BH08-21).
8. Vugs and cavities below the S-D contact, some of which show distinctive geopetal fabrics. The dolostone matrix are highly recrystalline. (Core 9. Consumers' Amoco 13076).
9. Typical dolomite-siliciclastic profile across the S-D contact. Thalassinoides commonly represent the top of the erosional surface of the Bass Islands Formation. A 1-cm thick unit below the blue mottled Thalassinoides is interpreted as joint infilled by Devonian sand. Oval and horizontally elongated breccias occur at the bottoms of the sinkholes and are overlain by glauconitic sandstone derived from the Springvale Member. (Core 16. OGS 82-2)
4.2.5 Karstification below the Silurian-Devonian boundary

The Silurian-Devonian unconformity in SW Ontario represents a major depositional hiatus (see Kobluk et al. 1977; Brunton and Dodge 2008). An extensive system of karst features, including solution-widened joints, solution-vugs, leached patches, sinkholes, collapse breccias, and in places filled with variable siliciclastic materials, is present below the disconformity surface in the upper Bass Islands/Bertie Formation (Kobluk et al. 1977; Hayens 1991). In Niagara area, the S-D unconformity is exposed in several quarries (see table 1.2), and this undulatory surface is regionally overlain by either the Devonian sandstone unit (Oriskany Formation or Springvale Member) or the Devonian carbonate rocks of the Bois Blanc Formation and/or Onondaga Formation.

Leaching and Vugs

The top 10–50 cm of the Bass Islands Formation exhibits pronounced leaching below the S-D contact in many cores. The leached interval is commonly pinkish to creamy in color, coarsely grained dolostone, and some contain quartz grains. Cements are bladed to bulky in thin section, and the quartz grains show some degree of abrasion. It is likely that the meteoric water has leached the top of the Bass Islands Formation during the Early Devonian and redistributed the Devonian siliciclastic sand grains into the leaching-enhanced porous space.

Vugs and small cavities (less than 10 cm in diameter) are extremely common near the contact, and occur through the top 3–5 m of the Bass Islands Formation. The vugs and cavities are irregular and either vertically or horizontally elongated. Dolomudstone or microbial-laminated dolomudstone plugged the bottom of the vugs/cavities, and the top is infilled by dolomite sparites (Fig 4.3-6). These vugs are connected by dissolution enhanced joints near the contact and become disconnected versus depth. A vertically trending system of vugs and leached interval has been
documented in the Niagara area (see Fig 3.1), and the vertically extensive patches are infilled by a mixture of dolomite, calcite cements, siliciclastic sandstone and glauconites. Diffused hematites are well present lined with the microfractures or groundwater flow zone.

The vugs and leached intervals represent an extensive freshwater solution and subaerial exposure during the latest Silurian to Early Devonian (Kobluk et al. 1977). The vugs and cavities may seem to have been formed randomly within 1 m below the contact, but other features, including the solution enlarged joints, oval and abraded breccias and Devonian sandstone infills, are found related to the vugs/small cavities zones (Fig 4.3-2). Vugs/small cavities are most abundant on either side of the solution enlarged joints, and decease in size and number away from the jointed or collapse brecciated intervals. Vugs are typically vertically elongated in the upper 30–50 cm and become preferentially parallel to bedding planes downwards. These features were unlikely formed in the phreatic zone because the vertically elongated vugs show dissolution by percolating meteoric water, and the abraded breccias indicate strong flow of near-surface fresh water. It cannot be ruled out, however, that these paleokarstic features were formed from phreatic cave structures that have been later elevated to the vadose zone during the tectonic uplift.

**In situ Breccias**

_In situ_ breccias are usually clast-supported, tens of centimetres below the contact, displaying small passageways and corroded clasts surrounded by wispy argillaceous seams (Fig 4.3-1 to 4.3-3). They show great variations in intercrystalline porosity that indicate incipient forms of the meteoric dissolution. However, these breccias are rarely collapse and maintain the same lithology with the context dolostone matrix. These breccias may have evolved into collapse breccias as the dissolution process continuous to dissolve the clasts and form a thin argillaceous layer as
undissolved residues on cave floor.

Joints

Joint systems are well developed below the contact, found in both subsurface cores (Fig 4.3-6) and outcrops in Niagara area (Fig 3.1C). Most of the joints are nonsystematic and penetrate the top 1.5 m of the Bass Islands Formation. These joints may be either closed, or enlarged open that are partly or completely infilled by dolomitic mud, euhedral dolomite minerals or Devonian siliciclastic sands. Celestite has also been found in the outcrop sections in Niagara area (personal communication with Derek Armstrong). Most of the joints display evidence of solution widening along length. Some joints have extended 1.5 m downward from the surface and are infilled by the erosional remnants (Fig 3.1C). A brecciated interval is commonly formed below the joint system (Fig 3.2B). Where Devonian siliciclastic sandstone is present (Fig 4.1E; 4.5H and 4.3-6), the coarse quartz grains could have infilled the joints that penetrated 1.2 m below the contact. Never have the joints been observed extended into the overlying Devonian units, which indicate all the joints were formed before the deposition of the Devonian units. In the discussion of water circulation in a karst terrane for the formation of solution-enlarged joint systems, Jennings (1971) and Sweetings (1973) noted that the solution-enlarged joints are formed by meteoric percolation. The joints can be infilled later by clay residues. Where the clay residues are removed into the joint systems, the karstic surface of the carbonates is overlain by younger deposits that are, in this study, the Devonian siliciclastic sands (Fig 4.1E; 4.5H and 4.3-6) or the Bois Blanc cherty carbonates (Fig 4.4-2).

Cave Deposits
Small caves or solution pits are found on top of the Bass Islands Formation in the Niagara area (Kobluk et al. 1977). Caves range from 1.5–3 m in diameter and are metres to tens of metres wide. In the outcrops, the solution pits are discontinuous with sharp or smooth edges with the surrounding Silurian dolostones. Modern analogues are the “kamenitzas” that represents the centimetres to metres wide, irregular karstic surface (Desrochers and James 1988). In the subsurface, solution pits or caves near the contact are hard to distinguish due to its large width, but speleothems are found 30 cm below the base of Bois Blanc Formation in core No. 5 (Fig 4.3-4). The speleothem consists of layers of meteoric dolomite cements that may have a stalactite or stalagmite origin. An angular joint has cut the layers of speleothem that have been infilled by younger dolomite cements. These solution features are also found together with all the other small-scaled dissolution features due probably to solution enlargement at the irregular erosional surface. These relative lowlands served to collect freshwater brought by rainfall and allow dissolution to enlarge the joint systems below. Unlike the caves in the lower Bass Islands Formation, the dolostone breccias above the cave floor show some degree of abrasion, which indicate a more robust dissolution water flow near the surface. These caves may have been reactive during the recent glacial period and played an important role in geoengineering hazards detect in Niagara area.

**Trace Fossils**

Pemberton et al. (1980) and Kobluk et al. (1977) reported Trypanites on the unconformity. The simple and unbranched borings in hard substrate with a single opening to the subsurface is easy to identify in field, which indicates an initial lithification during the S-D. In the subsurface, Trypanites have also been documented 3.5 m below the contact (Fig 4.3-9).
Another type of trace fossil, *Thalassinoides*, have been recognized right below the unconformity (Fig 4.1D; 4.3-9). It refers to the dichotomously or T-branched boxworks of burrow traces (Myrow 1995). *Thalassinoides* trace is expressed as irregular blue mottles within a creamy to light grey dolostone matrix, which are interconnected in a 3-D view. The burrowed unit is commonly 5–10 cm thick. Diverse marine organisms, including the sea anemone, fishes and decapod crustaceans, created *Thalassinoides* trace (Myrow 1995). It is likely that the intertidal crustaceans may have created these burrows during late Silurian to Early Devonian. The occurrence of this trace fossil usually represents the top of the Bass Islands Formation, and can be used as a regionally correlative marker bed in core observations. Locally, the *Thalassinoides* may have been overlain by the diagenetic collapse breccias or dolomite pebbles as cave/sinkhole infills on the paleokarstic surface of the Silurian rocks.

*Siliciclastic Joint and Sinkhole Infills*

The Devonian siliciclastic sandstone infills in the paleokarstic zone in the top part of the Silurian Bass Islands Formation consist of two types: joint infills and sinkhole infills. Joints are formed and enhanced by meteoric percolation near the exposed surface (Fig 4.3-5), and the migration of meteoric water may have distributed the Devonian siliciclastic sands along the joint walls. The joints that have been infilled by siliciclastic materials are thin (3-10 cm thick), and could reach as deep as more than 10 m below the contact (Summerson and Swann 1970). In core No. 16 (OGS 82-2) a 3-cm thick karstic conduit is infilled by Devonian siliciclastic sands 1.2 m below the contact (Fig 4.5H). In core 2 (U.S. Steel No.1) the sandstone unit occurs 30 cm below the top of the Bass Islands Formation (Fig 4.1D). These mixtures of Devonian sands within the Silurian dolostone matrix indicate a prolonged surface karst system during the Early Devonian.
Sinkholes were also present on the surface of the exposed Silurian carbonates. Dissolution and collapse of the carbonates below the contact has created many sinkholes on an irregular paleokarstic topographic surface (Fig 4.6-2). As the water table decreased, sinkhole floors begun to collapse and the sinkhole depth was increased. In the Early Devonian, the sinkholes may have been water-filled to facilitate a mixing of the descending Devonian siliciclastic sands with in situ Silurian carbonate breccias. The best example comes from Core No.7 (BH08-21), a 1.5-m thick siliciclastic sandstone of the Devonian Oriskany Formation has supported poorly graded, subangular dolostone breccias of various lithologies (Fig 4.3-7). The size of the breccias ranges from less than 1 cm to 5 cm, and lithologies varies from microbial-laminated dolomudstone, micritic dolomudstone to bluish mottled dolomudstone. The siliciclastic sandstones are interpreted as sinkhole infills based on their lack of bedding or other sedimentary structures. They were likely collapse breccias from the sinkhole walls because the subangular clasts are poorly bedded or graded. The surface water may have flushed the upper sinkhole sandstone fills and dissolved the surrounding sinkhole walls, so the collapse breccias fell into the sandstone matrix to form a sandstone matrix supported breccia fabrics. These dissolution and collapse processes have repeated several times during the Early Devonian exposure, thus forming a complex sinkhole system on the surface of the Bass Islands dolostones that could extend a few metres to tens of metres in depth. This irregular sinkhole systems also provide the accommodations for the Devonian siliciclastic sandstones to settle down, and could be one reason for the sporadic distribution of the Devonian sandstone units and their inconsistent thickness.
Figure 4.4: Silurian-Devonian contact in subsurface cores in SW Ontario

1. Contact of the lower Bass Islands Formation and the upper Springvale Member (Bois Blanc Formation) in area 3 (Core 3. Consumers’ Pan Am 13058). The lower dolostone is massive bedded, bleached with bivalve fragments. The S-D contact is a sharp and flat surface that is overlain by calcareous siliciclastic sandstone of Springvale Member. Brachiopod fragments are present 10 cm above the S-D contact.

2. Contact of the lower Bass Islands Formation and the upper Bois Blanc Formation in area 3 (Core 4. Consumers' Amoco 13061). A large dolostone nodule is present at the S-D contact. The cryptocrystalline dolostone in Bois Blanc Formation is very vuggy and cherty.

3. Contact of the lower Bass Islands Formation and the upper Oriskany Formation in area 3 (Core 11. OGS 82-3). The Oriskany quartz arenites are highly bioturbated and contain common brachiopod fragments.

4. Contact of the lower Bass Islands Formation and the upper Bois Blanc Formation in area 1 (Core 12. Imperial 805-Lyons No.1). The S-D contact is marked by a sharp lithologic change from the lower leached dolostone in Bass Islands Formation to the overlying cherty dolostone of Bois Blanc Formation.
**Figure 4.5** Lithofacies and karst features in Core 16 (OGS 82-2). The Bass Islands Formation is dominated by sabkha facies - comprising microbialites, argillaceous dolostone, anhydrite nodules and subaerial sedimentary structures. Shallowing upward cycles (3-5m thick) comprise subtidal massive dolostones through intertidal to supratidal microbialites possessing evaporite laths and anhydrite nodules.

A. Salina G shaly dolostone with common collapse structures - bedding is distorted due to sulphate (gypsum/anhydrite) dissolution.

B. Argillaceous dolostone. Note the dolostone nodule in the middle.

C. Anhydrite nodules grew within crinkly laminated algal laminites.

D. Dolomudstone with blue fenestrae mottles at base of Bass Islands Fm.

E. Brecciated dolostone.

F. Caliche deposits representing subaerial exposure.

G. Recrystallized dolostone, matrix is partially recrystallized.

H. Karstic conduit infilled by Devonian siliciclastic sands - 1.4 m beneath the S-D contact.

I. S-D erosional surface overlain by subrounded dolostone clasts - sinkhole infills.

*See* Fig 1.7 for legends.
4.3 Summary

The Bass Islands Formation represents the final stage of Silurian carbonate deposition in SW Ontario. The unconformity at the top of the formation marks a depositional hiatus of approximately two million years (Johnson et al. 1992). The prolonged erosional period has created a complex paleokarstic system within the Bass Islands Formation (Fig 4.7). Generalized karstification processes have been delineated in Fig 4.6. In general, three karstic intervals in the Salina G and the Bass Islands Formation can be correlated across SW Ontario from western Lake Erie to eastern Lake Huron into Michigan – the lower Salina G confined by two shale layers, the lower Bass Islands cave system and the top Bass Islands Formation below the contact (Fig 4.7). In its type section on Bass Islands, Ohio, the lower Put-in-Bay Member is characterized as very brecciated dolostones, which could be correlated to the lower karstic interval in lower-middle Bass Islands Formation. Upper karstic interval extends from the contact to metres below the top of the Bass Islands Formation. Dolines/sinkholes infilled by Devonian siliciclastic sandstones have been reported by Kerr (1950), Sparling (1970), Summerson and Swann (1970), Shaver et al. (1971) and Treesh (1972).

The lower Salina G Unit shows typical internal solution features that resulted from dissolution of previous existed evaporites. It is likely the two shale layers of Salina F and the basal Bass Islands have confined the Salina G as a separate aquifer that was hardly connected to other aquifer levels during Early Devonian (Fig 4.7). The regional extended cave system in the lower Bass Islands Formation is newly found, which usually caps the Cycle 1, and corresponds to a high kick in gamma-ray log. The upper paleo-karstic system occurs below the Silurian-Devonian contact has been regarded as formed during the regional uplift corresponding to the Salinic Disturbance in the earliest Devonian (Boucot 1968) when the Bass Islands Formation was exposed and top of which
has been eroded, thus a complex system of surface karstic features formed near the surface by possible a mixture of meteoric water percolation and by groundwater/marine water dissolution in the mixing zone. This protracted karstification has created an irregular surface that may have influenced the sporadic distribution of Lower Devonian siliciclastic sandstones of the Oriskany Formation (Summerson and Swann 1970; Brett and Ver Straeten 1994). The reactivation of the buried paleokarstic features in Ontario also plays an important role in bedrock aquifer evaluation.
The karstic features may have been formed by the dissolution of evaporites and the host dolostones both during and after the deposition of the Bass Islands Formation. The sequence of events envisioned for karstification processes are depicted in this diagram and described below.

1a-1d: Crinkly laminated microbialites acted as aquitards that confined the water flow in between to dissolve the anhydrite/gypsum nodules. The fluctuation of water table may also dissolve the underlying evaporites in the Salina Group strata and form the collapse structures and caves in the lower Bass Islands Formation.

2a-2d: Devonian sandstone can be found both above the S-D contact as sinkhole infills and in the karstic conduits below the S-D contact, suggesting an extended post-Oriskany karstification phase. Dissolution from the surface forms the vertical joints that have been infilled by Devonian sands or fallen clasts from surrounding walls in later stage.

3a-3d: Subaerial exposures during the deposition of the Bass Islands Formation have created several levels of paleokarstic intervals. The lowering of water table has enhanced the previously formed caves. Caves that developed along the water table are commonly oval to horizontally elongated, with matrix-supported cave infills. Phreatic caves are smaller and vertically elongated, with clast-supported infills.

4: Cave systems identified regionally in the lower Bass Islands Formation. Caves are usually 1-1.5m deep. The cave floor, cave infill and cave roof are readily recognizable in cored wells.

See Fig 1.7 for legends.
Evaporite-Dissolution Breccias

Epikarst (Sinkhole)

Sedimentary-Diagenetic Breccias

Buried Cave

a

b

c

d

Cave

1

2

3

4

Roof

Crackle Breccias

Collapse Breccias

Cave Floor
Figure 4.7 Paleokarstic profile of the Bass Islands Formation during the Early Devonian subaerial exposure period. Evaporite-dissolution breccias (marked as yellow) are confined to the units above the Salina G and the “False G”. A regionally extensive cave system is formed above the Akron Member that is traceable southwestward into the top of the “False G” unit. Depositional-early diagenetic breccias (marked as orange) are commonly found within the middle-upper Bass Islands and are only locally present. The prolonged exposure of the top of the Bass Islands Formation have created sinkholes and buried caves (marked as green) that are partly infilled by Devonian siliciclastic sandstones and dolomite breccias.
CHAPTER 5. The Silurian-Devonian Unconformities in Southwestern Ontario

5.1 Introduction

The compound Silurian-Devonian (S-D) boundary in Southern Ontario is marked by major depositional hiatus, which coincides with the boundary between the Tippecanoe and the Kaskaskia supersequences defined in North America (Sloss 1963; Johnson et al. 1992). The identification of these unconformities in Ontario is significant for establishing sequence stratigraphic and chronostratigraphic framework. In addition, features of the exposed surface may provide good insight into the active erosional processes during the hiatus. The upper Silurian carbonate rocks below the S-D boundary display significant erosional and paleo-karst features (Kobluk et al. 1977; Brunton and Dodge 2008; Sun et al. 2016). The differential erosion of the topmost Silurian rocks may have been related to the migration of the active forebulge (see Ettensohn and Brett 2002; Brett et al. 2004; Brintnell 2009; Brunton and Brintnell 2011; Brunton et al. 2012). Regional correlation of the unconformities helps identify eustatic lowstands and define paleotopographic highs and lows (Brett et al. 2000).

Kobluk et al. (1977) recognized two Pragian (before and after the Oriskany) erosional surfaces in the Niagara Peninsula. Brett et al. (2000) provided a summary of at least three Lower Devonian erosional surfaces in Niagara area and western New York. Sunderman (1988) described a similar siliciclastic-related S-D unconformity in Indiana. The extent of these significant unconformities has not been documented in the remainder of southern Ontario other than the Niagara Peninsula, in part, due to the lack of good outcrops of the S-D contact, presence of variably thick overburden, and nomenclatural differences at formation rank across national borders. This
chapter describes the S-D compound boundary in a peripheral forebulge region of SW Ontario, using selected cores (see Table 1.1) and outcrop sections (Table 1.2). The stratigraphic units that are related to these late Silurian-Early Devonian unconformities include, from bottom to top: the upper Silurian Bass Islands/Bertie formations, the Oriskany Formation, the Bois Blanc Formation (including its lower Springvale Member), and the Devonian Onondaga Formation in Niagara Region.

5.2 Stratigraphy

In this section, stratigraphic units related to the Silurian-Devonian unconformities are briefly discussed, from old to young. An overview of the stratigraphy is presented in Fig 1.1 (also refer to Chapter 2 for detailed description of the Silurian Bass Islands and Bertie formations).

5.2.1 Oriskany Formation

The Oriskany Formation is an Appalachian Basin unit and represents the oldest Devonian deposits in SW Ontario (Sanford 1969; Diecchio 1985). It is featured as coarse, well rounded, poorly sorted quartz arenite in Ontario (Reavely and Winder 1961). It is considered equivalent to the Oriskany Formation of New York State, western New Jersey, Pennsylvania, West Virginia, northeastern Kentucky and northeastern Ohio (Sanford 1969, 1993; Rickard 1975, 1984, 1985; Johnson et al. 1992; Scal 2002; Armstrong 2007; Kostelnik and Carter 2009; Klipp 2016) and Garden Island Formation of Michigan (Uyeno et al. 1982; Sanford 1993; Armstrong and Carter 2010). Previous studies identified the Oriskany Formation as an important gas producer (Cleaves 1939) in Pennsylvania and central New York (Kostelnik and Carter 2009). Vanuxem (1839) discovered a 6.1-m thick Oriskany sandstone overlying the Manlius Formation unconformably and established its type section at Oriskany Falls, Oneida County, New York. Within the Appalachian
Basin, it reaches a maximum thickness of 70-75 m in the Valley and Ridge Province and thins eastward (Bruner 1991). In western New York and Niagara Peninsula area, it is discontinuous and ultimately pinches out westward in the Hagarsville area (Uyeno et al. 1982). Maps showing the isopachs and distribution of the Oriskany Formation in eastern North America (Oliver et al. 1967, 1971; Boucot and Johnson 1967; Summerson and Swann 1970) indicate an outcrop area restricted to western New York and its adjacent Niagara Peninsula. An Oriskany-absent area to the east of the outcrop belt in northwestern Pennsylvania (Diecchio 1985).

In Ontario, Murray (1845) was the first to note a similar sandstone unit in the Niagara Peninsula. Later, Stauffer (1915) distinguished two sandstone units of different ages in the same area, and defined the younger unit as the Springvale Member at basal Bois Blanc Formation. Caley (1941, 1946) recognized a basal Devonian sandstone member from Fort Erie westward to Hagarsville, but was not able to distinguish sandstones of Springvale Member from Oriskany Formation. Best (1953) reported only one Oriskany occurrence 10 km east of Hagarsville. Despite its patchy distribution in Ontario and western New York, the Oriskany Formation may have a wider distribution in the subsurface but identification in well logs is difficult (Sanford 1968) due to its clean quartz sandstone lithology and high porosity, which has a flat signature in gamma ray and neutron logs and are hardly distinguished from the context carbonate units.

In Ontario, the Oriskany Formation consists of thick-bedded or massive, medium to coarse grained, loosely cemented quartz sandstones (Sanford 1967; Johnson et al. 1992; Armstrong and Carter 2010). Fragments of reworked Silurian dolostone are common at the base of the formation (Uyeno et al. 1982). Quartz grains are well-rounded to subangular and well sorted. Lenses of subangular sandstone conglomerates locally occur in outcrops (Derry Michener Booth et al. 1989), including Cayuga Quarry and Hagarsville Quarry (see Table 1.2).
Primary sedimentary structures have not been observed in either outcrops or cored wells. Bioturbated structures in thin section have been identified as burrow trace fossils. Megafossils in the Oriskany Formation are relatively common, mostly in moldic preservation (see faunal lists of Stauffer 1915 and Best 1953). Thick-shelled brachiopods, gastropods, and ostracods are present (Sanford 1967; Oliver and Hecht 1994). The Oriskany Formation is tentatively assigned to the Rosemarie faunal zone of Pragian age by Boucot and Johnson (1968) based on brachiopod studies. The more diagnostic brachiopods of Costispirifer, Rosemarie, Acrospirifer and Hipparionyx have been reported by Winder and Sanford (1972). Conodont study by Uyeno et al. (1982) shows that the Oriskany contains only fragments of long-ranging conodonts of little biostratigraphic utility, such as Icriodus sp. and Ozarkodina? sp. In both outcrops and subsurface, the Oriskany Formation overlies the Silurian dolostones of Bass Islands/Bertie Formation disconformably, marking the Silurian-Devonian unconformity (Uyeno et al. 1982). Micro-fractures, solution widened joints, cracks and pits infilled by Devonian sands have been reported by Kobluk et al. (1977) and have been observed in cored wells (see Chapter 2 in this thesis), as well as the trace fossil Trypanites by Pemberton et al. (1980).

In Ontario, the Oriskany Formation usually overlies the Silurian Bass Islands/Bertie Formation by an erosional surface that represents the latest Silurian-earliest Devonian hiatus. The contact is sharp and irregular, overlain by dripstones and basal cavity pebbles (Fig 4.4-7), which indicate that a paleokarst system was formed during the regional uplift event (Salinic Orogeny). In the cored wells, joints and channels are infilled with Oriskany sands are present (Fig 4.1E and 4.2H).

The depositional environment interpretation of the sandstones of Oriskany Formation has long been debated. The well-sorted, clean quartz minerals composition indicates a marine
depositional environment rather than fluvial. The sand provenance is uncertain. Baker (1983) suggested that the sands in the Oriskany Formation in the USA are considered to have been sourced from Appalachian Orogen. In western New York and the adjacent Niagara Peninsula, however, it may have an eolian origin from mixed source areas (Summerson and Swann 1970). Its constituent quartz sand grains may have derived from reworked Ordovician or even Cambrian siliciclastic materials blown from either the Appalachian Orogen, the northeastward Adirondacks or even the Wisconsin and/or Ozark domes to the west (Summerson and Swann 1970; Diecchio 1985). The uncertainty of clastic source material has hindered the interpretation of the Oriskany depositional environment.

The sandstones of the Oriskany Formation in northeastern USA is widely accepted as deposited in a marine environment (Seilacher 1968; Bruner 1991). Various models have been proposed, including beach shoal deposits (Swartz 1913), nearshore-shallow water deposits (Stowe 1938), and shallow to deeper subtidal deposits (Barrett and Isaacson 1977). Welsh and Bruner (1988) suggested tidal ridges and submarine dunes in a restricted shallow marine environment. There is a sandstone-free zone along central New York and northeastern Pennsylvania, separating the Oriskany sandstone in Ontario and western New York from the marine sandstones in other areas of the USA. This sandstone-free zone follows the axis of the Appalachian Basin (NE-SW), and was probably influenced by intermittent tectonic activities in the study area (Rosenfeld 1953; Bruner 1991). If so, this may indicate that the sandstones of Oriskany Formation in Ontario were formed in a periodic transition between shallow peritidal and subaerial environments. The sands may have been reworked and remobilized by wave action, and the occurrence of thick-shelled brachiopods *Cortispirifer arenosus* and *Rosemarie marylandica* is consistent with a high-energy environment (Klipp 2016).
5.2.2 Springvale Member (Bois Blanc Formation)

The Springvale Member occurs at the base of the Bois Blanc cherty limestone/dolostones (Stauffer 1913). In New York, it directly underlies the Onondaga Formation where the remainder of the Bois Blanc Formation is absent (Baker 1983). In its type locality near the village of Springvale in Haldimand County (Stauffer 1913), it consists of glauconitic and phosphatic, light brown-bluish or pinkish white, medium- to coarse- grained orthoquartzitic arenite with abundant wispy greenish and argillaceous microstylo-seams. Outcrops of this sandstone member are confined to the Niagara Peninsula, with the thickest accumulation of 3 m in Hagarsville area (Cowan 1977). It subcrops across southern Ontario, with the notable exception of Bruce and Huron Counties. Fauna is rich in Meristina and Amphigenia (Oliver 1966; Oliver and Sorauf 1981). Locally, hematite-stained clasts of dolomudstone and chert nodules are present in the upper Springvale Member. Sedimentary structures are readily observed in Core 5 Consumers’ Pan Am 13102 (Fig 4.4-3), including bioturbation and concentrated argillaceous seams. It can be differentiated from the underlying Oriskany Formation by presence of abundant phosphate and glauconite minerals, chert nodules and the higher calcite cements in cores. In geophysics logs, however, both of them have a flat and low gamma ray curve and high neutron log porosity, making it difficult to differentiate the log signature with the overlying carbonates of Bois Blanc Formation. Distribution of the Springvale Member is patchy and variable in thickness. In Core 13 (Argor 65-1) near Sarnia, for example, it is more than 10 m thick, but in Core 12 (Imperial 805 Lyons No.1) that is 20 km to the northeast, it is only 80 cm thick.

The upper contact with the overlying cherty carbonates of undifferentiated Bois Blanc Formation is sharp and flat. However, glauconite and phosphate minerals are observable at the base of the lowermost Bois Blanc cherty limestone/dolostone. Preferred depositional orientation
of quartz sands was not evident in cored wells with no observable primary sedimentary structures, which may indicate a possible quiet marine depositional environment. In thin sections, calcite cements are well developed around the quartz grains, in contrast to the quartz cements in the Oriskany Formation (Cowan 1977).

The lithologies of the Oriskany Formation and Springvale Member are very similar. Both consists of massive, white sandstones. Separation of these sandstone units is difficult unless with the preservation of fossils, glauconites/phosphates or petrographic microscope observation. The gamma ray signature of the Springvale Member is dependent on the content of carbonates and glauconite minerals. Where the radioactive glauconites are abundant, they create a positive shift in the gamma-ray response. Neutron porosity is comparable to that of the Bois Blanc Formation. Where the glauconite is absent, the Springvale Member is impossible to pick in geophysics logs.

5.2.3 Undifferentiated Bois Blanc Formation

In its type section on Bois Blanc Island in the straits of Mackinac, Michigan, the Bois Blanc Formation is characterized as grey-tan brown or bluish grey, finely crystalline limestones with abundant grey-white chert nodules (Sanford 1968). In the Niagara Peninsula and western New York State, overlying the Oriskany or Springvale sandstones, the Bois Blanc Formation consists of thin- to medium- bedded, microcrystalline, brownish grey dolostone or limestone with dark grey chert nodules, and contains large brachiopods, bryozoans, small rugose corals and rare tabulate corals. Further on the cratonic side westward of the Appalachian Foreland Basin, in Michigan and remainder of southern Ontario, it is lighter in colour and possesses more abundant light grey irregular-shaped chert nodules. The chert nodules are large (10-20 cm in diameter), and are commonly found within concentrated stylo-seams. Faunas are rich in brachiopods, bryozoans, gastropods and sponges. These filter feeder and/or detritus feeders were more adapted to elevated
nutrient levels. Rugose corals become abundant in the upper unit, indicating the depositional environment may have become shallower and warmer in the late stage of the Bois Blanc deposition. Both the faunal assemblage and the presence of large chert nodules indicate nutrient-rich waters where siliceous sponges could flourish. The dissolution of siliceous sponges is suspected to account for the source of silica to form such chert nodules (Musgrove 1983).

The Bois Blanc Formation is present as lenses in the Niagara Peninsula, and it varies in thickness from a thin veneer in Buffalo to a maximum of 5 m in Hagarsville and grades northwestward in Ontario to 30-50 m thick gray and tan dolomites in the Ingersoll-London areas and ultimately to 75-m thick, light grey limestones in Bruce and Huron counties. In the Sarnia-Woodstock area, the Bois Blanc Formation has an argillaceous and glauconitic basal interval where the Springvale Member sandstone is absent. Due to the pervasive chertification of the Bois Blanc Formation, detailed correlations are difficult to make from the Appalachian Basin into the Michigan Basin (Cassa and Kissling 1982). Syn-depositional sedimentary structures are obscured by diagenetic chert nodules and dolomitization, thus the lithologic character of the carbonate rocks and the proportion of marine fauna and diagenetic alteration fabrics are potentially helpful.

In cores, the Bois Blanc can be easily distinguished from the underlying sandstone units by its lack siliciclastic components, abundant chert nodules and a very high porosity. Geophysics logs for the Bois Blanc Formation display a flat low gamma signature. Its upper contact with the Amherstburg Formation, however, is an unreliable top-pick in Ontario (Armstrong and Carter 2010). Where the sandstone of the Sylvania Formation is present below the Amherstburg Formation in Essex County (e.g. Core 22 Cansalt DDH 87-3), the basal contact is readily recognizable by a gradational change from sandy dolostone at the top of the Bois Blanc Formation into the loosely packed and friable sandstone of the Sylvania Formation. Where the Sylvania
Formation is absent, the contact seems gradational. Only in cored wells can this contact be identified. There is a lithologic change from the massive bedded, light grey, sparsely fossiliferous and very cherty limestone/dolostone of the Bois Blanc Formation into the light brown, bituminous, cherty and fossiliferous Amherstburg Formation. The fauna, however, has changed significantly. Bois Blanc Formation features common benthic filter feeders and the Amherstburg Formation contains a more diverse fauna of abundant corals, crinoids and brachiopods.

5.3 Unconformities

At least two unconformities are recognizable at the contact between the Lower Devonian Bois Blanc Formation and the upper Silurian Bass Islands/Bertie formations in the Niagara Peninsula, while only one unconformity surface can be identified in the remainder of southern Ontario by subsurface bedrock mapping. In the Niagara Peninsula, this surface is overlain by Early Devonian quartz-rich sandstones of the Oriskany Formation. The sub-Bois Blanc/Springvale unconformity rests directly on the Oriskany Formation sandstone or directly on the Silurian dolostones (Fig 5.1 and 5.2) that probably removed all the Oriskany and part of the upper Bass Islands and Bertie formations. In southwest of southern Ontario, from Sarnia across Chatham to southern Lake Erie, the Springvale sandstone overlies the S-D unconformity surface but the clean Oriskany sands are not present. In the northwest of southern Ontario, from Bruce County to Huron County, only one unconformity is evident between the Bass Islands dolostone and Bois Blanc dolostone/limestone.

5.3.1 Sub-Oriskany Unconformity

Evidence of this unconformity can be found in all cores in the Niagara Peninsula where Oriskany quartz arenites overlie the Silurian dolomite by a sharp and irregular unconformity surface (Fig 5.1). Kobluk et al. (1977), Ciurca (1982, 1990, 1994) and Ciurca and Hamell (1994)
noted that this unconformity is complex, regionally variable, and probably a combination of more than two erosional unconformity-forming episodes (see also Brett et al. 2000). At the Nelson Quarry, the S-D contact represents a slightly irregular surface of the Oriskany Formation sitting disconformably on the Silurian Akron Member of the Bertie Formation (Fig 5.2A). Locally angular dolomite clasts of Bass Islands/Bertie Formation occur in the sandstones of Oriskany Formation and solution-widened joints are infilled by clean quartz arenite (Fig 5.1 and 5.2). Under the S-D unconformable surface, boring trace fossils (*Trypanites*) and microfractures are common. Vugs commonly plugged by secondary dolomites (refer to Kobluk et al. 1977).

In cores, however, typical solution-widened joints or *Trypanites* trace fossils are not readily recognizable, but *Thalassinoides* are well developed, and disarticulated brachiopods are mixed with the clean quartz sands that overlie this contact (Fig 5.1A). Dark grey, irregular karstic surfaces are locally found under the S-D contact (Fig 5.1D). Presence of clasts of poorly sorted, subangular dolomudstone that are cemented by the quartz sandstone usually marks this unconformity. For example, in Core 7 BH08-21, a solution-widened cavity is found at the unconformity surface with clasts of subangular dolomudstone surrounded by quartz grains at the base.

### 5.3.2 Sub-Bois Blanc/Springvale Unconformity

From Port Colborne to Hagarsville, outcrops of the Springvale Member occur directly on either the Oriskany Formation or the Silurian Bass Islands/Bertie formations (Fig 5.1 B-C). Kobluk et al. (1977) described the development of neptunian-like dikes, solution pits, vugs and multiple joint sets containing Oriskany Formation sandstone. In the Hagarsville Quarry, this irregular unconformity is associated with phosphatized clasts of the massive sucrosic dolomudstone of the Bass Islands Formation, abundant glauconite, concentrated argillaceous seams, and calcareous cements (Fig 5.1A). Disarticulated brachiopods are commonly found overlying the basal contact.
and become larger and more common in the upper unit in cores (Fig 5.1B). In the Nelson Quarry (Fig 5.2B), the contact displays low relief steps of 1-m long joints and borings.

In the subsurface of southwest southern Ontario, the unconformity surface is irregular. Glauconite and sand grains commonly infill the large cavities of the uppermost Silurian dolostones. In Core 12 (Imperial No.809 – W.J. Mawson No.1), for example, the sandstone in the Springvale Member is dark green and very glauconitic. Overlying the S-D unconformity, irregular clasts of Silurian dolomites are mixed with the glauconitic sandstone, which grades upwards into cherty, sandy dolostone (Fig 5.1F). In Core 13 (Argor 65-1), this unconformity overlies a 40-cm thick karstic brecciated zone that consists of subangular dolomite clasts mixed with glauconitic sandstone (Fig 4.4-5). In Core 16 (OGS 82-2), a 30-cm thick, thinly bedded sandstone unit overlies the S-D contact, and is overlain by a 50-cm thick unit of sandy glauconitic dolostone. Vugs and cavities are very well developed under this unconformity (Fig 5.1D). Irregular vugs are highlighted by meteoric leaching and the microfractures are commonly plugged by sparry dolomite. Oil stains are also pervasive (Fig 5.1B). Episodes of irregular karstic surfaces that consist of undissolved clay remnants, are found in Core 22 (Cansalt DDH 87-3), indicating a complex, multi-phased erosional history.

5.3.3 Amalgamated Sub-Oriskany and Sub-Bois Blanc/Springvale Unconformity

In Bruce and Huron counties, the Bois Blanc carbonates rest directly on the upper Silurian strata by an amalgamated erosional surface (Fig 5.1E). At the present, however, it is not clear whether this surface represents a non-depositional hiatus or that the Lochkovian–upper Emsian deposits were subsequently eroded. Westward, across the forebulge from Niagara Region, the basal bluish grey, microcrystalline dolostone or limestone of Bois Blanc overlies the uppermost Silurian dolostone by a flat, 3–5 cm thick interval with closely spaced argillaceous seams (Fig 5.1H). The
uppermost Silurian dolomudstone is microfractured and infilled by secondary dolomite (Fig 5.1E), with common irregular vugs highlighted by leaching, within which sand grains are lacking. Where the massive light grey dolomudstone is present under the unconformity, the porosity is enhanced to form a sucrosic texture (e.g., Core 1. Domtar Goderich S.T. #1, Fig 5.1E). Locally the uppermost Silurian dolostones are hydrocarbon stained. In this area, geophysics profiles show little change across this unconformity because of the similar lithology of the lower Bois Blanc Formation and the upper Bass Islands Formation and possess a high porosity.

5.4 Summary and Conclusions

The S-D unconformity represents a complex and regionally extensive paleokarstification surface in southern Ontario. The upper Silurian succession represents restricted marine to shallow marine conditions (e.g., microbial laminites and intertidal massive dolomudstone in arid subtropical environment, see Chapter 2). Locally on the farfield side of the Algonquin Arch the peritidal lithofacies changed upwards into more normal marine/lagoonal facies where evaporite deposits decreased, indicating a regional sea level rise.

The pre-Oriskany erosional events may have eroded substantial upper Silurian strata compared to those uppermost Silurian units in northern Appalachian Basin (Laporte 1969, 1971) and southern North America craton (Broadhead et al. 1988), and represented an initial stage of the deposition of upper Lower Devonian sands (upper Pragian). The sands are clean, well sorted and coarsely sized, and are not continuous on regional basis. Commonly, the sands are best preserved in paleokarst sinkholes or solution-enhanced joints. The Springvale glauconitic and phosphatic quartz-rich sandstone mixed with carbonate minerals create a calcareous sandstone zone that marks a second erosional phase. The post-Oriskany sea level fluctuation may have eroded all
Oriskany sands where the Springvale Member overlies the dolostone of the Bass Islands Formation. Although leached, vugs and microfractures infilled by secondary dolomite/calcite are very common in the uppermost Bass Islands and Bertie formations. Paleo-soils are absent due to either non-deposition or were removed by the chemical dissolution/paleokarstification. The spotty distribution of the two sandstone units may indicate an irregular seafloor and intermittently subaerial surface (e.g., Karrow 1973) perhaps due to the dissolution of the underlying Silurian evaporites (Sanford and Brady 1955; Summerson and Swann 1970). The distribution of the well sorted and coarsely sized sandstone offers evidence for an initial transport of the sands by wind (Summerson and Swann 1970), and the successive marine invasions immobilized it and mixed it with fossils. Along the eastern Michigan Basin margin, the basal sandy and glauconitic deposits of the Bois Blanc Formation may be coeval with the Springvale sandstone in the Appalachian Basin in late Early Devonian. The overlying cherty, fossiliferous carbonates of the Bois Blanc were probably deposited on a marine shelf or ramp with diverse assemblages of abundant brachiopods, bryozoans, sponges, gastropods and small rugose corals, representing the return to marine carbonate deposition scenario through Early to Middle Devonian.
Figure 5.1. Examples of the S-D contacts in selected cores. Each core length is 35 cm long and the top is to the left. Scale bar = 2cm.

A) Contact of the Oriskany Formation quartz arenite (Devonian) and its underlying Bass Islands Formation (Silurian). The Oriskany Formation sandstone contains common fragments of brachiopods and overlies the sucrosic dolostone of Bass Islands Formation unconformably. Contact surface is 1cm thick with common clays and quartz grains cemented by dolostone (Core 2. U.S. Steel No.1).

B) Contact of the Bois Blanc Formation shaly limestone (Devonian) and its underlying Springvale Member (Devonian). The lower sandstone of Springvale Member is phosphatic and glauconitic, containing fossils of brachiopods (*Amphigenia*). Porosity of the sandstone is very high – note the very oil stained interval to the right (Core 4. Consumers’ Amoco 13061).

C) Contact of the pinkish phosphatic sandstone of Springvale Member and its underlying dolostone of Bass Islands Formation. The blue mottled dolostone is microfractured and the unplugged vugs are commonly highlighted by leaching (Core 10. Consumers’ Pan Am 13057).

D) Dolostone of Bass Islands Formation. Large cavity that has been infilled by secondary sparry dolomite 10 cm below the contact of the overlying Bois Blanc Formation. Irregular karstic surface can be seen 5 cm below the cavity (Core 19. Consumers’ 33409).

E) Dolostone of Bass Islands Formation under the contact with the overlying Bois Blanc Formation. Microfractures and irregular vugs are very common due to subaerial exposure and leaching, which are partly plugged by secondary dolomite (Core 1. Domtar Goderich S.T. #1).

F) Contact of the glauconitic dolostone of Bois Blanc Formation and the faintly laminated dolostone of Bass Islands Formation. The contact surface is irregular, and compacted dark green glauconite minerals infilled the surface. The lower Bois Blanc Formation dolostone is still glauconitic, with irregular light grey chert nodules to the left (Core 8. Imperial No.809 – W.J. Mawson No.1).

G) Contact of dolostone of Bois Blanc Formation with the massive dolomudstone of Bass Islands Formation. Contact surface is sharp and flat which is overlain by an irregular chert nodule unit at the base of Bois Blanc Formation. Dolostones above the chert nodule unit is slightly glauconitic. No vugs or fractures or karstic features can be seen in the top of the Bass Islands Formation (Core 14. Imperial 805 – Lyons No.1).

H) Contact of the dolostone of Bois Blanc Formation with the Bass Islands Formation. Two irregular karstic surfaces can be seen below the contact. Dissolution surfaces are commonly infilled by light grey laminated dolostone (Core 22. Cansalt DDH 87-3)
Figure 5.2. Examples of S-D unconformities (significant time breaks) in outcrops.

1) Contact of Bois Blanc Formation (Devonian) and the Bass Islands and Bertie formations (Silurian) in Hagarsville Quarry. The bottom 2 m is the buff to light blue mottled Akron Member of the Bertie Formation, which is overlain by 4.5-m thick, medium to massive bedded dolomudstone. The S-D contact is irregular with common dissolution pits that are infilled by argillaceous dolostone of Bois Blanc Formation. The lower 1.5 m of the dolomudstone of Bois Blanc Formation is thinly bedded and very argillaceous with common glauconite and phosphate. The top 1.1 m is the undifferentiated cherty Bois Blanc Formation dolostone. The Oriskany Formation is not present in this quarry. (Hammer scale is 33 cm in length).

2) Contact of the Devonian undifferentiated Bois Blanc Formation/Springvale Member/sandstone of Oriskany Formation in the Nelson Quarry. The lower 2 m is the Oriskany Formation quartz arenite with sparsely scattered oil stains. Note the lighter grey lenses or nodules of sandstone with more quartz grains in the medium grey sandstone matrix. Contact of Springvale Member and Oriskany Formation is horizontal. A joint is present in the right middle part of photo that is partly infilled by the Springvale Member glauconitic and phosphatic sandstones. The Springvale Member is only 20 cm thick and is overlain by the thinly to medium bedded dolostone of Bois Blanc Formation with common light grey chert nodules. (Hammer scale is 33 cm in length).
CHAPTER 6. Lithofacies and Stratigraphy of the Devonian Onondaga Formation and Dundee Formation in Niagara area, southwestern Ontario

6.1 Introduction

The Onondaga Formation is an upper Lower Devonian–lower Middle Devonian unit overlying the pre-Onondaga Formation siliciclastic sandstones (e.g. Oriskany Formation and Springvale Member in Bois Blanc Formation) in the Appalachian Foreland Basin (Oliver 1954, 1956; Brett and Ver Straeten 1994; Brett et al. 2011). In the New York State and the Niagara Peninsula of Ontario, it is characterized by brownish grey to medium dark grey, cherty and fossiliferous limestones (Oliver 1954, 1956; Cassa and Kissling 1982). The Onondaga Formation is well studied in New York State, but less well known in Ontario. Currently, the name “Onondaga Formation” is not used in the Oil, Gas and Salt Resources Library, Ontario Petroleum Data System (OPDS) database because it is regarded to be equivalent to part of the Detroit River Group, as a result of different stratigraphic nomenclature from that of the United States (Rickard 1975, 1984; Johnson et al. 1992; Armstrong and Carter 2010). The Onondaga Formation was previously considered to be stratigraphically equivalent to the Amherstburg Formation (Oliver 1954, 1956; Fagerstrom 1961, 1966) and is so identified in all Ontario petroleum well cards (e.g. Armstrong and Carter 2010). Another reason for its not having been incorporated in the OPDS database is that it has the same stratigraphic position with the Detroit River Group in SW Ontario. Both the Detroit River Group and the Onondaga Formation in Ontario are underlain by the Bois Blanc Formation and overlain by the Dundee Formation (Fig 1.1). Outcrops of the Onondaga Formation are found in the Niagara region of Ontario (Fig 6.1 and 6.2), but possess a different fauna and lithology from those of the Amherstburg and Lucas formations (Detroit River Group) in other areas.
Figure 6.1 Well and outcrop locations of the Onondaga Formation in SW Ontario. Please refer to table 1.1 for well numbers and well names. The subsurface Onondaga lithofacies grades into the Detroit River Group lithofacies from Core 10 to 11 and the transitional zone is supposed to locate between Port Burwell and Port Bruce.
Figure 6.2 Paleozoic bedrock geology of southern Ontario. The Onondaga Formation subcrops in the Niagara area as well as beneath western Lake Erie. It grades into Amherstburg and Lucas formations westward in northern Norfolk County. The eastern edge of the Onondaga extends through Lake Erie. *(modified from OGSR Library, 2017).*

of Ontario (Oliver 1954, 1956; Fagerstrom 1966) (Fig 6.3). Therefore, it is necessary to separate the Onondaga Formation from the Detroit River Group in SW Ontario, and reassess the stratigraphic relationships of the carbonate-dominated succession across SW Ontario, eastern Michigan and northern Ohio.
In general, the stacked carbonate succession of the Onondaga Formation displays low radioactivity on gamma-ray logs, even though the vertical lithofacies vary regionally. The fossiliferous and dark grey cherty limestones of the Onondaga Formation represent an open marine depositional environment, which differs from the more restricted lagoonal or cratonic deposits of the Amherstburg Formation and the sabkha-dominated deposits of the overlying Lucas Formation. It is speculated that the movement of the forebulge region has largely controlled the lithofacies variations between the Detroit River Group in Michigan Basin and the Onondaga Formation in Appalachian Basin (see Root and Onasch 1999; Ver Straeten and Brett 2000; Ettensohn and Brett 2002; Brett et al. 2004; Britnell 2009; Brunton and Britnell 2011; Brunton et al. 2012).

Conodont biozones of the Detroit River Group are poorly constrained in Ontario (Telford et al. 1977; Uyeno et al. 1982), making it difficult to correlate the succession of Onondaga Formation from Ontario into New York and Ohio. Recently, based on regional sequence stratigraphy, Brett et al. (2011) suggested that the Onondaga Formation is entirely younger than the Detroit River Group and is only equivalent to the Anderdon Member of the upper Lucas Formation (see Fig 6.3). The Anderdon Member is the uppermost pure limestone or sandy limestone unit of Detroit River Group (see Birchard et al. 2004 for details) and yields a good conodont recovery (Uyeno et al. 1982). This revision has left more questions regarding the Lower–Middle Devonian stratigraphic correlations in SW Ontario. This chapter is a first attempt to refine the lithofacies distribution and subsurface stratigraphy of the Onondaga Formation in SW Ontario, and to provide a conceptual stratigraphic framework in the Niagara Peninsula region that extends into western New York State.

The study area, where the Onondaga Formation subcrops, is situated from Fort Erie westward into Port Burwell along Lake Erie (Fig 6.1 and 6.2). Tectonically, it is on the farfield side of the Appalachian Foreland Basin, in a peripheral bulge (forebulge) setting that is generally known as the
Findlay-Algonquin Arch (Fig 1.1 and 6.1; see Ettensohn and Brett 1998; Root and Onasch 1999). The depositional area of the Onondaga Formation is to the east of this forebulge region along and south of the Onondaga Escarpment in the northwestern region of the Appalachian Basin. The axis of the Appalachian basin is northeast-southwest trending in central New York (Ettensohn 2008). During the Onondaga deposition period, a forebulge region was inferred east of the Niagara area on the northwestern margin of the Appalachian Foreland Basin (Cassa and Kissling 1982; Ver Straeten and Brett 2000). Therefore, the migration of the forebulge may have played an important role in the lithofacies distribution of the Onondaga Formation and the Detroit River Group in Ontario and its adjacent New York State.

6.2 Previous Studies of the Onondaga Formation

In 1828, Eaton was the first to report the “Corniferous Limerock” that was equivalent to what is now called the Onondaga Formation in New York. Hall (1841) described this unit as a cherty dolomitic limestone unit in Onondaga County, New York. Originally it was placed as equivalent to the Corniferous Limestone of the upper Helderberg Group (Lochkovian, Early Devonian) by Boyd (1881, in Dennison 1961). Willard (1936) separated the Helderberg Group from the Onondaga Formation, and Woodward (1943) formally introduced the term “Onondaga Formation”. Later revisions made by Cooper et al. (1942) and Dutro (1981) placed the Onondaga Formation at the top of Emsian (late Early Devonian), overlying the Schoharie Formation in eastern New York. Other names including Ulsterian (Schuchert 1943) have been proposed to correlate Schoharie and Onondaga Formations but now have been abandoned.

Currently, Oliver’s (1954; 1956) chronological scheme of the Onondaga Formation is widely accepted and used in the reports published by New York Geologic Survey. In eastern New York,
Oliver (1954) identified four members, which in ascending order, including: Edgecliff, Nedrow, Moorehouse and Seneca (Fig 6.4). Though the Onondaga Formation is not fully exposed in Hall’s (1839) first described type site, Oliver (1954) has correlated it to the upper Edgecliff Member through to the middle Seneca Member in his subdivision. In western New York and the Niagara region, a fifth member, Clarence, has replaced the stratigraphic position of the Nedrow, between the Edgecliff and the Moorehouse (Oliver 1966; Cassa and Kissling 1982). It differs from the Nedrow Member by not containing any shale layers (Oliver 1956).

In Ontario, the Onondaga Formation crops out in a narrow belt along the north shore of Lake Erie and continues to Hagarsville where it underlies glacial sediments (see Fig 6.1. Telford and Tarrant 1975a, 1975b; older terms are summarized in Ehlers and Stumm 1951; Stumm et al. 1956). South of the outcrop belt, it is present in the subsurface and grades into the Amherstburg Formation cherty limestones and the overlying Anderdon Member of the Lucas Formation along the western shore of Lake Erie (see Fig 6.2; Birchard et al. 2004). In Ontario, only the Edgecliff, Clarence and lower Moorehouse members are present; the upper Moorehouse and the Seneca Member are considered as replaced by the Dundee Formation in the Niagara region evidenced by a lithological transition zone of Seneca dark grey limestone to Dundee Formation light brown, cherty encrinites east of Buffalo (Oliver 1966a; Sanford 1967).
Figure 6.3 Stratigraphic diagram with local names shown and correlated for regions of southeastern Michigan, northern Ohio, SW Ontario, western and central New York. Group names are in upper cases; formation names in normal text. Different views on Devonian strata correlation in Lake Erie region are listed in last two columns (Rickard 1975, 1984; Brett et al. 2000, 2011). Abbreviations: Mbr = member.
6.2.1 Lithology

In general, the Onondaga Formation in western New York and Ontario consists of medium to dark grey, cherty and fossiliferous limestone (Rickard 1981, 1984; Johnson et al. 1992). The lithology of each member is not disputed. The Edgecliff Member is rich in fossils that have made up 50% of the rocks. Diagnostic fossils include abundant large crinoid columnals, colonial rugose corals (*Acinophyllum* and *Synaptophyllum*), solitary rugose corals (*Heterophrentis*, *Heliophyllum* and *Cystiphyllum*), tabulate corals (*Cladopra* and *Thamnopora*) and fenestral bryozoans. Bioherms contain tabulate corals (*Cladopora*) and colonial rugose corals (*Acinophyllum*) are locally developed in western New York and Niagara area, Ontario (see Wolosz 1982, 1991, 1992, 1995a, 1995b). They are usually 6–15 m thick, up to 200 m in diameter (Wolosz and Paquette 1988). Other fossils in bioherms include crinoids, solitary rugose corals, brachiopods, trilobites, gastropods and bryozoans (Oliver 1954, 1956). Centimetre-thick concentrations of fossiliferous grainstone beds are locally developed and cemented by sparry calcite (Lindholm 1969).

The Nedrow Member is an argillaceous and clayey, fossiliferous limestone characterized by platyceratid gastropods in central and eastern New York (Oliver 1954). In western New York and Ontario, the Clarence Member has the same stratigraphic position and is characterized by dark grey lime-mudstone with very abundant black chert nodules and rare fossils. The black chert nodules may make up more than 50% of the overall rock volumes of the Clarence Member. The Moorehouse Member overlies the Nedrow Member, identifiable by a pair of shale beds in central New York (Brett and Ver Straeten 1994). In Ontario, however, the shale beds are absent, making it difficult to correlate to its type section in New York. The Moorehouse Member consists of fossiliferous dark grey limestones that have similar lithofacies to the Edgecliff Member, though the rugose coral fauna is different (Fig 6.4). Colonial rugose corals (e.g. *Acinophyllum*) are not present, and the solitary rugose corals are dominated by *Eridophyllum* and *Heliophyllum*. Bioherm
facies, however, is not present in the Moorehouse Member.

The overlying Seneca member is similar in lithology with the Moorehouse Member, and the contact is marked by a regionally traceable potassium-rich ash bed, termed the Tioga Bentonite in New York (Dennison and Textoris 1978; Tucker et al. 1998), which can be traced into Indiana (Droste and Vitaliano 1973). This ash bed has been reported 9-12 m above the base of Dundee Formation in Ontario by Sanford (1967) and Winder and Sanford (1972), though no type section or reference core was given.

The thickness of the Onondaga Formation varies from 46 m (Feldman 1985; Rickard 1989; Brett and Ver Straeten 1994) in Allegheny province, eastern New York, to about 25 m in the Ohio Valley, central New York, and to nearly 44 m in Buffalo to Syracuse, western New York (Prosser et al. 1913; Wolosz and Paquette 1988; Brett and Ver Straeten 1994). In the subsurface of Ontario, it is 12-18 m thick. The Onondaga Formation in central New York thickens both westward and eastward towards the Appalachian Basin margins (Rickard 1981; Feldman 1985; Brett 2011). The thickness of the individual members varies extensively according to various authors due to different interpretations of lithological facies, lithological similarity between Edgecliff and Moorehouse, and local absence of the Clarence-Nedrow Member. In Ontario and western New York, the occurrence of bioherms in Edgecliff Member adds to the discrepancy of the thickness of each member. Even in the same area, different authors reported different thicknesses (see Oliver 1954; 1956; Dennison 1961; Cassa and Kisling 1982; Wolosz 1991, 1992, 1995a, 1995b; Feldman 1994). Wolosz and Paquette (1988) have summarized that the Edgecliff pinnacle reefs are present in eastern New York, and the biothermal Edgecliff Member is thinner in western New York, though the Clarence-Nedrow, Moorehouse and Seneca are thicker. Generally, where the Edgecliff bioherm facies is present, its thickness may reach up to 30 m in western New York and Ontario, and the
overlying Clarence-Nedrow Member is thinner or absent (Crowley and Poore 1974).

The Onondaga Formation in Ontario and western New York has been interpreted as carbonate deposits on an easterly dipping ramp on the farfield side of Appalachian Foreland Basin (Dennison 1975; Dennison and Textoris 1978; Kissling and Moshier 1981; Dennison 1986; Dorobek and Read 1986; Brett and ver Straeten 1994). The four readily recognizable members were described in detail by Oliver (1954) in the type area, south of Syracuse, based on lithofacies, faunas and marker beds of shales and potassium-rich ash beds. The Niagara area was interpreted as a shallow inner to middle ramp, shallow open marine depositional environment, where no shale beds were found (Cassa and Kissling 1982).

### 6.2.2 Biostratigraphy

The biostratigraphy of the Onondaga Formation in New York and adjacent Ontario (Fig. 6.4) has been studied for more than 60 years. Klapper (1971, 1981) recognized two Conodont zones in the Onondaga, a lower “*Polygnathus patulus*” zone and an upper “*Polygnathus costatus costatus*” zone. The former has an unknown base through up to the Nedrow or lower Moorehouse beds that may cross the Emsian–Eifelian (Early Devonian-Middle Devonian) contact, and the latter has a base from the middle Moorehouse into Seneca (Oliver and Klapper 1981). The conodont biozones are poorly constrained in its presumably equivalent Detroit River Group, SW Ontario, although the Anderdon Member of the Lucas Formation is rich in conodonts and has been assigned to the *P. patulus* zone that is correlative to the lower Onondaga Formation. The lower part of Dundee Formation has been placed in the *P. costatus costatus* zone (Uyeno et al. 1982; Sparling 1985). Therefore, the Dundee Formation is biotratigraphically correlative with the upper Moorehouse and Seneca members (Telford et al. 1977; Oliver and Klapper 1981; Uyeno et al. 1982).
Figure 6.4 The biozones and common faunal components in the Onondaga Formation in New York and Ontario. The Emsian and Eifelian boundary is placed in the lower Nedrow Member in New York based upon conodonts (Oliver and Klapper 1981). Data are collected from Oliver (1954; 1956), Sparling (1970), House (1981), Oliver and Sorauf (1981), Uyeno et al. (1982), Lindermann and Feldman (1987), Kirchgasser and Oliver (1993). (ON=Ontario; NY=New York). Thickness of each member is not to scale. See Fig 1.7 for legends.

Based on brachiopods (Oliver 1966a, 1966b, 1967), Dutro (1981) and Lindermann and Feldmann (1987) have assigned the Bois Blanc Formation to the small Ampheginia zone and the Edgecliff, Nedrow and lower Moorehouse to the large Ampheginia zone. Fagerstrom (1966) discovered the last occurrence of the orthide brachiopod Dalejina in both the uppermost Amherstburg and the upper Edgecliff units and suggested they are correlative. Five trilobite species have been identified by Ludvigsen (1987) from the Formosa reef (uppermost Amherstburg
Formati, and four of them occur in the Edgecliff member. Oliver and Sorauf (1981) have summarized the rugose coral biostratigraphy and recognized two assemblage zones within the Onondaga Formation, the *Acinophyllum-Synaptophyllum* Zone in the Edgecliff Member and the *Eridophyllum-Heliophyllum* Zone in the Nedrow and Moorehouse Members. Goniatite biostratigraphy is poorly defined for the Onondaga Formation, with only one species found that has been placed in House’s (1981) Fauna #3.

**6.2.3 Regional Correlation**

The correlation of the Onondaga Formation into the Michigan Basin in Michigan and Ohio has long been problematic (e.g. Sanford 1967; Rickard 1975, 1984; Uyeno et al. 1982; Sparling 1985; Brett et al. 2011) because of a lack of field outcrops displaying the contact relations of the two depositional units in SW Ontario.

Though has long been regarded as equivalent to the Amherstburg Formation in Ontario (Oliver 1966; Sanford 1967; Rickard 1975, 1984; Uyeno et al. 1982), minor revisions have been made on the correlation scheme of the Onondaga Formation based on the conodont data collected in Michigan and Ohio by Sparling (1985). He suggested that the Dundee Formation from Ingersoll to Goderich in Ontario and in the Michigan Basin is younger than the Dundee Formation that overlies the lower Moorehouse Member in Niagara area, evidenced by the highest occurrence of “*P. costatus costatus*” above the erosional surface at the top of Dundee Formation. In the Michigan Basin, however, the highest occurrence of “*P. costatus costatus*” is within the lower Dundee Formation. Therefore, he suggested that in Niagara area and western New York, the Dundee Formation and its equivalent upper Moorehouse and Seneca members are older than the Dundee Formation in the Michigan Basin, and could thus be correlated to the upper Lucas Formation in Michigan. Other evidence comes from the correlation of the Tioga ash beds occurred in the
Onondaga Formation to the Kawkawlin bentonite beds that occur in the upper Lucas Formation in Michigan (Baltrusaitis 1974). In this correlation scheme, the Dundee Formation in the Michigan Basin should be the chronological equivalent to the base of Marcellus Shale in New York (Sparling 1985). It is also possible that because the Dundee Formation in Niagara area was located on the forebulge region, short-lived uplift of the forebulge region could lead to differential erosion of the upper Dundee Formation and the section above the “P. costatus costatus” zone is all gone.

Brett et al. (2011) proposed alternative explanations for such lateral lithological patterns, suggesting that the top of the undifferentiated Lucas Formation should be placed below the base of Onondaga Formation, and the Detroit River Group regarded as Early Devonian deposits that are older than the Onondaga Formation, thus being equivalent to the Bois Blanc and Schoharie Formations in New York. They have also inferred that there was a major non-depositional gap between the Lucas Formation and Dundee Formation, during which the Onondaga Formation was deposited in New York. This correlation contradicts almost all the previous biostratigraphic studies (e.g. Landes 1951; Fagerstrom 1966, 1971, 1982; Linsley and Kesling 1982; Ludvigsen 1986). Therefore, a revision on the stratigraphic position of the Onondaga Formation is necessary to achieve in Ontario.

Chapter 6-8 attempts to: 1) clarify the lithofacies and member subdivisions of the Onondaga Formation in Ontario; 2) place the lithostratigraphy into a sequence stratigraphic framework; and 3) establish a subsurface correlation of the Onondaga Formation to other Devonian units in Ontario.
**Figure 6.5** Stratigraphy and geophysics characteristics of the Onondaga Formation. Example of lithologies and geophysics data is taken from Core 4 (Consumers’ Amoco 13061). See Fig 1.7 for legends.
6.4.3 Lithofacies Descriptions and Interpretations of the Onondaga Formation

In this study, seven lithofacies have been identified within the Onondaga Formation (LF2-LF8 in Table 4.1), representing the lower Edgecliff Member to lower Moorehouse Member overlain by the Dundee Formation (LF9-LF13) in Brant, Welland, Haldimand and Norfolk Counties (Fig 6.2). In general, the cherty limestone of the Onondaga Formation in Ontario represents shallower water lithofacies than the basinal to deeper subtidal facies described from central and eastern New York (Oliver 1956).

In Ontario, all three members of the Onondaga Formation (Edgecliff, Clarence and Moorehouse) display low radioactivity and a flat character in gamma-ray profiles, except for a positive shift of gamma-ray log and a decrease in neutron porosity in the lower-middle Edgecliff Member contact. The neutron logs, however, may be used for regional stratigraphic correlations (Fig 6.5). The two sand units beneath the Bois Blanc and/or Onondaga formations consist of clean quartz arenite that display a lower gamma-ray response than the carbonates. This gamma inflection may represent the base of a transgressive unit (Fig 6.4). The Onondaga Formation is recognizable in neutron log profiles by having a higher neutron porosity than the underlying Bois Blanc Formation and the overlying Dundee Formation. A gradational drop of neutron porosity usually marks the upper contact of the Onondaga Formation with the Dundee Formation. The Dundee Formation maintains a flat gamma-ray and neutron geophysical character with a sharp increase of radioactivity at the upper contact of the shales in Marcellus Formation.
Table 6.1 Lithofacies types of the Bois Blanc, Onondaga and Dundee formations in Niagara area. LF1 belongs to the Bois Blanc Formation, LF2-8 to the Onondaga Formation, and LF9 to LF13 to the Dundee Formation. (see Fig 6.5 for stratigraphy and Fig 6.6-6.10 for core photos).

<table>
<thead>
<tr>
<th>Lithofacies (LF) Description</th>
<th>Depositional Environment</th>
<th>Formational Occurrence</th>
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<tr>
<td>LF1. Cherty Brachiopod Dolomudstone:</td>
<td>Compact, massive bedded, light to medium gray dolomudstone with packstone-grainstone pulses (Fig 6.6 LF1). Brachiopods (small <em>Ampheginia</em>) and bryozoans are common, and rugose corals rare. Displacive, whitish grey, rimmed chert nodules are very common, making up more than 50% of the whole formation. Fossils are also found in the irregular, rimmed chert nodules, including brachiopod fragments, trilobites, ostracods, gastropods (<em>Platyceras</em>) and rugose corals (<em>Metriophyllum, Edaphophyllum</em> and <em>Acrophyllum</em>). Argillaceous, horsetail stylo-seams commonly occur around the chert nodules. A few centimetres long vertical fractures are locally infilled by secondary dolomites. Brecciated chert nodules are present, truncated by secondary dolomites near the top of this lithofacies. Centimetre-thick fossiliferous packstone intervals are locally present, with very common fragmented rugose corals and articulate brachiopods.</td>
<td>Open marine, continental shelf, near the storm wave base. Packstone/rudstone pulses may represent tempestite intercalations. Brecciated chert nodules may indicate that a paleokarstic system was developed at the depositional hiatus prior to the deposition of the Onondaga Formation.</td>
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LF2. Quartz Arenite: | White to pinkish or greenish grey, very glaucocitic and phosphatic quartz sandstone with various intervals fragments of solitary rugose corals (e.g. *Metriophyllum* and *Cystiphymum*), brachiopods (large *Ampheginia*) and gastropods (*Platyceras*) (Fig 6.6-LF2; Fig 6.7-1 and 4). The fauna assemblages are different from those in the Springvale Member of the Bois Blanc Formation and the Oriskany Formation that are rich in brachiopod fragments. In thin sections, the quartz grains show degrees of frost, abrasion, and secondary cements are ring are observed. This sandstone unit has a sporadic distribution in Niagara area. Its thickness ranges from absent to nearly 12 m (the thickest is found in Core 2. U.S. Steel No.1). It combines with the Springvale Member where the Bois Blanc Formation is absent in western-central New York. | Reworked windblown sands infilled the paleo-topographic lows as sand ridges on the karstic domain of the top of Bois Blanc Formation. |Unnamed sandstone unit below the Onondaga Formation |
LF3. Cherty Coral-Brachiopod Wackestone-Packstone: Medium brown, thin-medium bedded, slightly bituminous and locally argillaceous, bioturbated wackestone-packstone. Fossils include abundant tabulate corals (*Favosites*), solitary rugose corals, brachiopods, crinoids and fenestral bryozoans (Fig 6.6 LF3a and 3b; Fig 6.7-2 to 3 and 5 to 8). Gastropods and trilobites are rare. Storm beds that are 10-15 cm thick, consisted of coral and brachiopod fragments are locally present. Chert nodules are light grey to medium brown, either replacive or displacive. Burrow trace fossils and bioturbation is pervasive. Burrows are partly plugged by integrated lime muds. Hardgrounds are present near the top of this lithofacies and marks the gradational contact with the overlying crinoidal-packstone.

LF4. Crinoid Wackestone-Packstone: Dark grey, thinly-medium bedded and locally nodular bedded, argillaceous wackestone-packstone with abundant crinoid columnals and ossicles (Fig 6.6-LF4; Fig 6.8-1 to 5). Size of the crinoid debris ranges from 1 mm to more than 5 mm. Brachiopods and rugose corals are rare. Chert nodules are less common than LF3, making up less than 10% of the facies. Wispy stylo-seams are common. Thickness is 2-3.5 m. Basal contact is marked by the occurrence of large crinoid ossicles and a color change from the light brown LF3 to the dark grey LF4.

LF5. Coral-Crinoid Packstone-Wackestone: Dark grey, thin to medium bedded, biostromal packstone to wackestone with fossiliferous floatstone interbeds. Fossils are dominated by solitary and colonial rugose corals (*Acinophyllum* and *Synaptophyllum*) (Fig 6.6-LF5; Fig 6.8-6 and 7). Crinoids, tabulate corals, brachiopods are very common. Chert nodules are very rare. Beddings are locally pseudonodular surrounded by argillaceous stylo-seams. Biohermal lithofacies have been reported along the Onondaga outcrop belt, which consists of colonial rugose corals (*Acinophyllum*) and tabulate corals (*Cladopora* and *Thamnopora*). Thickness ranges from less than 2 m to 12 m. Basal contact with the underlying crinoidal packstone is gradational.
**LF6. Cherty Mudstone:** Massive bedded, light grey to dark grey, finely crystalline mudstone with abundant black chert nodules (Fig 6.6 LF6). Black chert nodules have made up 50% to more than 90% of this lithofacies. Fossils are rare to common, including platycerid gastropods, triolobites and small rugose corals. Thin beds of fragmented rugose corals and brachiopods packstone are locally present as intercalations. Trace fossils of *zoophycos* are present. Thickness ranges from 0.8 m to 2.5 m. This lithofacies represents the typical Clarence Member, and the contact with the underlying Edgecliff Member and the overlying Moorehouse Member are gradational. LF6 interfingers with the fossiliferous packstone-floatstone (LF5) are common in the basal gradational contact zone.

**LF7. Coral-Crinoid-Stromatoporoid Wackestone:** Dark grey, medium bedded, bioturbated, very fossiliferous wackestone with a finely crystalline matrix (Fig 6.6-LF7; Fig 6.9.1 to 3). Lithofacies is similar to LF5, but contains different fauna. Rugose corals (*Heliophyllum, Cystiphyllum and Eridophyllum*) and crinoid debris are very abundant. Brachiopods and tabulate corals (*Favosites, Cladopora and Thamnopora*) are common but colonial rugose corals are not present. Laminar stromatoporoids are present. Chert nodules are light brown, irregular and replacive, making up less than 10% of the whole lithofacies. Thickness ranges from 2.5m to more than 8m. Together with the intercalations of LF6 and LF8 it forms a shoaling cycle near the base and the top of the lower Moorehouse Member.

**LF8. Fossiliferous Grainstone:** LF8 has two end members. LF8a: 10-15 cm thick, bluish grey, brachiopods and crinoids dominated grainstone (Fig 6.6 LF8; Fig 6.904). Rugose corals are rare. Skeletal grains are fragmented and tightly compacted, forming a crinoid shoal facies that represents the shallowest deposits occurring in middle Moorehouse Member. Thin beds of LF8a occur repeatedly below the upper contact with the crinoidal limestone or sandy limestone base of the Dundee Formation. LF8b: medium bedded, light brown, coarsely crystalline grainstone with very abundant rugose corals and brachiopod fragments. The matrix is highly recrystalline by leaching and coarse marine cements are present. Intercrystalline porosity is common, the overall permeability, however, is low due to recrystallization and cementation. Basal contact with the dark grey LF7 is sharp.
**LF9. Bioturbated Cherty Crinoidal-Pelloital Mudstone-Wackestone:** Light to medium brown, slightly argillaceous, crinoid rich limestone with common white chert nodules and argillaceous stylo-seams (Fig 6.10-1 and 2). This lithofacies may have a sandy base overlying the fossiliferous Moorehouse Member by an unconformity. Quartz grains are sparsely dispersed in the wackestone matrix. Skeletal grains include common crinoids and dark grey pellets. Brachiopods, trilobites, tabulate corals (*Thamnopora*) and solitary rugose corals are rare. Matrix is highly bioturbated. Irregular bluish grey bioturbation marks are common. Chert nodules are light grey, rimmed and irregular in form. Chert nodule beds seem to be parallel to the beddings planes. The chert nodules tend to form along the fossiliferous intervals and replace the original fossil skeletal grains. Sparsely distributed small irregular chert nodules was interpreted as replacement of the burrow trace fossils. Argillaceous wispy stylo-seams are very common around the chert nodules. Quartz grains may occur at the base of this lithofacies, in the form of well rounded, frosted and subangular grains. Dolomite rhombs (sucrosic dolomites) are present. Packstone/grainstone pulses are found as intercalations within this lithofacies, with common crinoid and brachiopods fragments that have been interpreted as tempestites. The crinoid proportion increases towards the top of this lithofacies. Burrows are irregular, subhorizontal near the top of this lithofacies. This lithofacies is 5-8 m thick. The basal contact with the Moorehouse Member is sharp and readily recognizable.

**LF10. Fenestral-Pelloital Grainstone:** Buff to light grey, massive bedded, pelloital grainstone with high porosity (Fig 6.10-3). Peloids are fine grained carbonate sands with fragments of brachiopods, bivalves, crinoids, corals and recrystalline gastropods. Peloids are spherical to subspherical in form with uniform size. Skeletal size varies. Fenestral porosity is very common that have been enhanced by dissolution. Interparticle and intraparticle porosity also feature this lithofacies.
**LF11 Crinoid-Coral Wackestone-Packstone:** light to medium brown, slightly bituminous packstone with common light grey chert nodules and argillaceous stylo-seams (Fig 6.10-4 and 5). Crinoids are abundant and tabulate corals (*Favosites*) are common. Rugose corals, bryozoans and brachiopods are less common. Interbeds of 3-5 cm thick, fossiliferous (crinoids and bryozoans) grainstone pulses are present. Chert nodules are in the same form with those found in LF9. Matrix is slightly argillaceous by the fact that concentrated argillaceous stylo-seams are commonly developed around chert nodules. Chert nodules make up 10% of this lithofacies. Near the top of this lithofacies, both the chert nodules and fossil contents decrease. Horizontal burrows occur near the top. This lithofacies is 3-5m thick. The basal contact with the underlying LF9 is gradational, and can be only recognized as a change of fauna contents.

**LF12. Bituminous Mudstone:** Medium to dark brown, thinly to medium bedded mudstone with very rare fossils (Fig 6.10-6). Solitary rugose corals, brachiopods and gastropods occur sparsely within the mudstone matrix. Individual mudstone beds are separated by wavy carbonaceous seams. Chert nodules are in the same form of those in LF9, but are less common and only make up less than 5% of the lithofacies. Bluish burrow marks are well developed. This lithofacies occurs in the middle of Dundee Formation and its thickness ranges from 3-6 m. The basal contact is marked by a sharp surface with the underlying fossiliferous LF10. The possible Tioga ash bed occurs at the top of this lithofacies that marks the upper contact with LF12. The potassium rich ash bed is dark grey, 5 cm thick, rich in biotite with white clay clasts. This marker bed represents the contact of Moorehouse Member and Seneca Member in New York.

**LF13. Crinoid-Brachiopod Wackestone.** Medium grey, medium to massive bedded, coarsely grained wackestone with very abundant crinoids and brachiopods (Fig 6.10-7 and 8). Other fossils include fenestral bryozoans and trilobites. Grainstone interbeds occur locally. Firmgrounds are well developed (Birchard 1993) that consists of bluish grey wackestone overlain by bottom dwelling crinoid holdfast and inarticulate brachiopods. Bluish horizontal burrow traces are common. The matrix is highly bioturbated. This lithofacies contain no chert nodules and is less argillaceous. Thickness of this lithofacies is 4.5-9 m. This lithofacies is commonly absent in the Niagara area due to recent erosion. Where it is preserved, it is overlain by the black, brachiopod rich shale of Marcellus formation by a sharp contact.

 Protected shallow marine, inner-middle ramp above the storm wave base. Dundee Formation

 Quiet water, possibly dysaerobic, restricted lagoon. Middle Dundee Formation

 Outer ramp, deep water deposits below the storm wave base. Upper Dundee Formation
Figure 6.6 Selective lithofacies of Bois Blanc and Onondaga formations. Each sample is 15 cm in length and the top is to the left. Samples are taken from Core 4. Consumers’ Amoco 13061. LF1 (198.10-198.25 m): Very cherty dolomudstone of the Bois Blanc Fm. The slightly argillaceous dolostone is confined by two irregular and displacive chert nodules above and below. Note the brachiopod fragments found both in chert nodules and in the dolomudstone matrix and common fractures in the lower chert nodule. LF2 (189.75-189.9 m): Quartz arenite of the unnamed sandstone below the Onondaga Formation. Glauconite minerals are common in the lower 5 cm and phosphates common in the upper 3 cm. Fragments of brachiopods and rugose corals are of Onondaga age. Argillaceous seams are present and the sandstone is locally oil stained. LF3a (186.05-186.20 m): Coral-Crinoid grainstone of lower Edgecliff. Coarsely-grained coral fragments are overlain by crinoid-articulate brachiopod grainstone that indicate a possible high energy depositional environment. LF3b (184.05-184.20 m): Cherty Coral-Crinoid Wackestone. Typical lithofacies of the lower Edgecliff (Amherstburg-like). Note the chertified rugose coral to the left overlies the unaffected calcareous tabulate coral. Porosity is intercrystalline and partly plugged by secondary calcite. LF4 (176.90-177.05 m): Crinoid Packstone. Non-cherty packstone dominated by crinoid ossicles that represent the middle Edgecliff. LF5 (173.75-173.90 m): Coral-Crinoid Packstone. Common biostromal facies that represents the upper Edgecliff. Fauna are dominantly *Synaptophyllum* and *Cystiphyllum*. LF6 (170.50-170.65 m): Cherty Mudstone. Sparsely fossiliferous Clarence Member characterized by abundant black chert nodules. Note the black chert nodules surrounded by light grey chert nodules, suggesting at least two phases of chert growth or diagenetic reaction rims. LF7 (160.00-160.15 m): Coral-Crinoid Wackestone. Lower Moorehouse Member that resembles the biostromal Edgecliff, but different in fauna. Corals of *Cladopora, Favosites, Heliophyllum* and the occurrence of lamellar stromatoporoids indicate a more robust fauna community of Moorehouse age. LF8 (156.05-156.20 m): Crinoid-Brachiopod Grainstone. Crinoid ossicles are very abundant with rare articulate brachiopods. Matrix is coarsely crystalline, indicating a shallow water deposits that’s possibly formed by the winnowing of storms or waves. The contact of Moorehouse Member (Onondaga Formation) and the Dundee Formation is marked by the argillaceous limestone at the top 3 cm. LF9 (155.90-160.05 m): Cherty, argillaceous mudstone of the Dundee Formation Chert nodules are light grey in color and commonly replacive. *Br* = Brachiopod, *Cr* = Crinoid, *RC* = Rugose Coral, *TC* = Tabulate Coral, *By* = Bryozoan, *St* = Stromatoporoid, *Ch* = Chert, *BC* = Black Chert, *Ph* = Phosphate, *Gl* = Glaucnite, *AS* = Argillaceous Seams.
Figure 6.7 Lithofacies of LF2 to LF3.

1. LF2 (Core 3. Consumers' Pan Am 13058; 172.85m). Large tabulate corals grew on top of the clean quartz arenite. Brachiopod fragments are commonly present in the lower sandstone unit. The intercrystalline porosity within the tabulate corals are partly plugged by the pure quartz grains.

2. LF3 (Core 4. Consumers' Amoco 13061; 195.55m). Dark grey, argillaceous limestone with gastropod fossils. The chert nodules are large, with a sharp edge against the limestone matrix. Concentrated carbonaceous seams are present near the top.

3. LF3 (Core 2. U.S Steel. No.1; 134.0m) Top of the lower Edgecliff Member. Note the brecciated chert nodules with fissures.

4. LF2 (Core 2. U.S Steel. No.1; 126.35m). Unnamed sandstone below the Onondaga Formation. The Onondaga-aged Synaptophyllum are sparsely distributed within the quartz arenite.

5. LF3 (Core 2. U.S Steel. No.1; 121.54m). Tabulate corals grow on top of a grainstone interval that have been replaced by chert nodules.

6. LF3 (Core 5. Consumers' Amoco 13102; 167.9m). Abundant corals near the top of the lower Edgecliff Member. Some rugose corals are fragmented.

7. LF3 (Core 3. Consumers' Pan Am 13058; 161.7m). Bluish grey bioturbation marks in LF3. Matrix is pelloital.

8. LF3 (Core 2. U.S Steel. No.1; 119.0). Fragmented rugose corals near the top of the lower Edgecliff Member, indicating a high-energy environment near the end of the lower Edgecliff deposition.
Figure 6.8 Lithofacies of LF2 to LF3.

1. LF3/LF4 (Core 10. Consumers' Pan Am 13057; 141.2m). Contact of the lower and middle Edgecliff Member. The top of the lower Edgecliff Member is marked by the tan grainstone with very common corals and crinoids. Basal middle Edgecliff Member consists of large amount of crinoid debris.

2. LF3/LF4 (Core 2. U.S Steel. No.1; 114.06.0m). Contact of the lower and middle Edgecliff Member. The top of the lower Edgecliff Member is marked by brecciated chert nodules with bryozoans. Abundant crinoids characterize the base of the middle Edgecliff Member.

3. LF4 (Core 10. Consumers' Pan Am 13057; 132.0m). Typical lithology of the LF4. Abundant solitary rugose corals (Synaptophyllum) overlie the basal colonial rugose corals (Acinophyllum)

4. LF4 (Core 3. Consumers' Pan Am 13058; 129.3m). Biostromal facies of the LF4. Fossils include lenticular stromatoporoids, rugose corals, tabulate corals (Cladopora) and brachiopods.

5. LF4 (Core 5. Consumers' Amoco 13102; 146.5m). Oil stained tabulate corals near the top of the upper Edgecliff Member. Good intercrystalline porosity in tabulate corals forms a potential oil reservoir.

6. LF4/5 (Core 3. Consumers' Pan Am 13058; 138.15m). Contact of the Edgecliff and Clarence Members. Fossils become very rare above the dark brown chert nodule.

7. LF5 (Core 7. BH08-21; 32m). Typical lithology of the Clarence Member. Fossils are very rare and black chert nodules are pervasive.
Figure 6.9 Lithofacies of LF7 to LF8 and Tioga ash beds.

1. LF7 (Core 10. Consumers' Pan Am 13057; 114.15m). Very abundant rugose corals (Eridophyllum) in Moorehouse Member. Cladopora is present near the base.
2. LF7/LF8 (Core 2. US Steel. No.1; 94m). Contact of the Moorehouse Member and the overlying Dundee Formation. Lenticular stromatoporoids, bryozoans, Cladopora and crinoids are common at the top of the Moorehouse Member. The basal Dundee Formation is slightly sandy.
3. LF7/LF9 (Core 5. Consumers' Amoco 13102; 138.67m). Contact of the Moorehouse Member and the overlying Dundee Formation. The basal Dundee Formation is very sandy. Quartz grains are observable within a limestone matrix.
4. LF8 (Core 2. US Steel. No.1; 111.15m). Very abundant Amphipora that form up a back-reefal or lagoonal facies in Moorehouse Member.
5. Tioga ash bed (?) (Core 5. Consumers' Amoco 13102; 107.55m). A possible Tioga ash bed is present 24 m above the base of the Dundee Formation.
6. Tioga ash bed (Core 11. OGS 82-3; 114.35m). Tioga ash bed is present 22 m above the base of the Dundee Formation. Biotites are rich and small light colored illites are common.
7. Tioga ash bed (Core 9. Consumers’ Amoco 13076; 78.45m). Tioga ash bed is present 20 m above the base of the Dundee Formation. Dolomite clasts float in the lower and upper interval of the ash bed.
Figure 6.10 Lithofacies of LF9 to LF13.

1. LF9 (Core 7. BH08-21; 24.4m). Contact of the Dundee Formation and the underlying Moorehouse Member. Tabulate corals occur at the top of the Moorehouse Member. The intercrystalline porosity is partly infiltrated by sand grains.

2. LF9 (Core 5. Consumers' Amoco 13102; 137.37m). Grainstone intercalations within the LF9 in lower Dundee Formation. Very common crinoids, brachiopods and pellets characterize the grainstone intervals.

3. LF10 (Core 4. Consumers' Amoco 13061; 128.25m). Fenestral limestone with highly bioturbated matrix and tempstite beds. Tempstites consist of corals and brachiopod fragments. Moldic porosity is present in the fenestral limestone matrix.

4. LF11 (Core 2. U.S Steel. No.1; 88.6m). Coral-crinoid wackestone to mudstone. Matrix is highly bioturbated. Chert nodules are replacive, irregular in form.

5. LF11 (Core 5. Consumers' Amoco 13102; 129.1m). Coral-crinoid wackestone of LF11. Spore cases of *Tasmanites* are also present.

6. LF12 (Core 5. Consumers' Amoco 13102; 118.4m). Typical lithology of the LF12. Very cherty mudstone with very rare fossils. Brachiopod fragments are present. Rimmed chert nodules are dark color, irregular in form.

7. LF13 (Core 2. U.S Steel. No.1; 84.7m). Brachiopod-crinoid wackestone to mudstone. Note the geopetal infill of the brachiopod in the lower section. Crinoids accumulate above a bioturbated mudstone matrix that represents a possible firmground surface.

8. LF13 (Core 5. Consumers' Amoco 13102; 103.6m). Brachiopod-crinoid wackestone. Limestone matrix is highly bioturbated. Peaked stylolites are present near the top section.
CHAPTER 7. Stratigraphy and Paleoecology of the Onondaga Formation and Dundee Formation in Niagara area, Southwestern Ontario

The Bois Blanc Formation, Onondaga Formation and Dundee Formation in Niagara area, Ontario, consist of more than 45 m of stacked carbonate deposits (Armstrong 2017). In Ontario, the basal contact of the Onondaga Formation is marked by a regional unconformity above either the Lower Devonian Bois Blanc Formation, Oriskany Formation or the Silurian Bass Islands Formation in Ontario and western New York (Oliver 1966; Cassa and Kissling 1982; Brett and Ver Straeten 1994). Six types of basal Devonian sequences have been recognized by Oliver (1966a, 1967), in which two siliciclastic sandstone units are present at the base of the Bois Blanc and the Onondaga formations. The sandstone units below the Bois Blanc Formation have been discussed in Chapter 5, this thesis. The sandstone unit below the Onondaga Formation is newly found in Ontario, which represents another unconformity below the Onondaga Formation. This thickness of sandstone unit varies from absent to 15 m in the subsurface of Ontario (Oliver 1966a). The sandstone unit in the lower Bois Blanc Formation is termed as the Springvale Member in Ontario or Springvale Sandstone in the United States (Boucot and Johnson 1968). The sandstone unit below the Onondaga Formation in Ontario, however, has long been considered the Springvale Member interfingering with or overlying the Bois Blanc Formation carbonates (e.g. Armstrong and Carter 2010). In New York State, the sandstone unit below the Onondaga Formation is also referred to as the Springvale Sandstone where the Onondaga Formation overlies the Silurian units directly and the Bois Bland Formation is absent (Cassa and Kissling 1982). Because the name “Springvale Member/Sandstone” has been applied to two sandstone horizons of arguably different age, this terminology is confusing and in need of revision. In this chapter, the Springvale Member refers to the sandstone below the Bois Blanc Formation and the “unnamed sandstone below Onondaga” to
the sandstone unit below the Onondaga Formation.

In Ontario, the upper contact of the Onondaga Formation with the overlying sandy or argillaceous and cherty limestone of the Dundee Formation is marked by a regional irregular disconformity. The Dundee Formation is in turn overlain by the brachiopod-rich shale of Marcellus Formation. In central New York, there is a regional shallowing event marked by the occurrence of crinoid grainstones that separates the lower and upper Moorehouse Member. Brett et al. (2011) regarded the top of the grainstone as a third-order sequence boundary that may be correlative with the boundary of the Anderdon Member and the Dundee Formation in Ontario (Fig 6.3). The upper Moorehouse and the Seneca Member of the Onondaga Formation directly underlies the Marcellus Shale (Oliver 1956) in New York.

Detailed stratigraphic description and refinement of each unit of the Onondaga Formation are present in the following section, based on Oliver’s subdivision scheme (1954, 1956).

7.1 Bois Blanc Formation in Niagara area

The Bois Blanc Formation comprises the basal Springvale Member and undifferentiated limestones or dolostones (Oliver 1966). In the subsurface, it is overlain by the unnamed sandstone below the Onondaga Formation and/or the limestones of the Onondaga Formation directly. Where the unnamed Onondaga sandstone is absent, a 1–3.5 m thick glauconitic and argillaceous unit is present at the top of Bois Blanc Formation carbonates, overlying a disconformity of unknown duration (e.g. Ridgemount Quarry South and in Core 10. Consumers’ Pan Am 13057 at the depth of 162.2–163.2 m).

The Bois Blanc Formation carbonates in the Niagara Peninsula, identified as LF1 (Table 4.1). The Bois Blanc Formation fauna, which is dominated by sparsely scattered small brachiopods
(small *Ampheginia* and *Leptaena* sp.), bryozoans, rare small rugose corals, trilobites and gastropods, differs from that in the overlying Onondaga that consists of abundant rugose corals, crinoids and large brachiopods (large *Amphigenia*). The Bois Blanc Formation contains a similar brachiopod assemblage to that of the Schoharie Formation in eastern New York and thus was considered of the same age and correlative across the Appalachian Basin (Caley and Liberty 1957; Oliver 1967; Liberty and Bolton 1971; Brett and Ver Straeten 1994), though detailed correlation of each bed is impossible. Distinctive fauna of *Aemulophyllum exiguum* (Billings) and *Acrophyllum oneidaense* (Billing) have been reported by Oliver (1954, 1966a, 1966b, 1967, 1968) and have been recognized in cored wells in the Niagara area (Fig 6.7-1). This species has also been reported from Schoharie Formation in eastern New York (Oliver 1958). Platycirid gastropods are present and usually chertified (Fig 6.7-2).

Chert nodules are uncommon in New York (Oliver 1966b; Brett and Ver Straeten 1994) but pervasive in Ontario. The thickness of the Bois Blanc Formation varies from absent west of Buffalo to 3 m in Port Colborne and at least 5 m in Hagarsville (Armstrong and Carter 2010). In the subsurface bedrock in the Niagara area, its thickness is 3-8 m and the top usually represents a regional cut-down associated with the pre-Onondaga erosion caused by local migration of the forebulge (Ver Straeten and Brett 2000). One exception is in the subsurface of Port Burwell area where Bois Blanc Formation carbonates are not present and the basal Onondaga Formation sandstone overlies the Silurian Bass Islands Formation directly by a 12-m thick siliciclastic unit (e.g. Core 5. Consumers’ Amoco 13102).

In both outcrops and subcrops the Bois Blanc Formation limestones form a distinct unit and can be readily recognized by its fauna and light grey, compact and finely crystalline lithology. The upper contact is marked by an abrupt lithologic and faunal change from light grey cherty limestone
to either sandstone units below Onondaga Formation or a medium brown, cherty coral-crinoidal wackestone of the lower Edgecliff Member. Locally, a shale layer occurs 1 m below the upper contact in outcrops (e.g. Dunnville Quarry) that may represent paleokarstic dissolution residues. In geophysics logs, this contact is difficult to place because of the flat gamma ray response in both the Bois Blanc and overlying Onondaga carbonates, except where the paleokarstic profiles are well developed with large amount of undissolved residues that has a high response in the gamma ray log. Because of the lateral discontinuity of the undissolved residue layer it cannot be used as a regional correlation marker. The neutron log, however, is somewhat useful to characterize the upper contact, corresponding to an increase of neutron porosity in the overlying sandstone unit (Fig 6.5).

**Depositional History and Paleoecology**

The Bois Blanc Formation has been dated as Emsian (Early Devonian) based upon brachiopod assemblage (Boucot and Johnson 1968), though the large tabulate corals yield an Emsian-Eifelian age (Early to Middle Devonian; Oliver 1966a). The Bois Blanc Formation has a poor conodont recovery, with only one species of *Icriodus latericrescens robustus* reported by Uyeno et al. (1982). This is a long time-range conodont species and has little biostratigraphic application. Klapper and Ziegler (1967) recovered *I. huddlei* in western New York, which is regarded as Lower Devonian age (Klapper and Ziegler 1979), and placed the Bois Blanc Formation in Emsian stage of Early Devonian.

During the Emsian (407-397 Ma), the Niagara area and western New York were possibly located in an epeiric sea environment that was restricted by the forebulge region to the east in central New York (Ver Straeten and Brett 2000). The deposition of the Bois Blanc Formation in
Niagara area began after a major regression that deposited the widely-distributed siliciclastic sandstone of the Springvale Member (Summerson and Swann 1970; Davis 2017). The presence of glauconite, phosphates and reworked argillaceous sand grains indicate a prolonged period of non-deposition prior to the deposition of Bois Blanc Formation. The subsequent transgression possibly moved from Laurentia craton interior easterly into the Niagara area. The Bois Blanc Formation in Niagara area is thought to represent the eastern feather edge of its distribution (Oliver 1960). The 3-8 m thick Bois Blanc Formation in Niagara area is much thinner than in the rest of Ontario. For example, in Goderich, the Bois Blanc Formation in Core No.1 Goderich S.T. #1 has reached nearly 35 m. The thinning of the Bois Blanc Formation is due to either regional erosion in the forebulge region across Lake Erie during Emsian or the deposition of the lower Bois Blanc-aged Springvale Member sandstone has reinforced an unwelcome environment for the carbonate build-ups in Niagara area. The depositional environment interpretation of the Bois Blanc Formation is difficult to achieve because the original beddings, sedimentary structures and porosity have all been destroyed by the growth of replacive chert nodules. Only the fauna assemblages and various forms of stylolites can be used as correlative markers.

The brachiopods in the Bois Blanc Formation in Niagara area include *Amphigenia*, *Acrospirifer duodenaria* (Hall), *Coelospira camilla* (Hall) and *Centronella glansfagea* (Hall) (Uyeno et al. 1982). Recognition of the fragmented brachiopods are difficult down to species level in cored wells, but the *Amphigenia* is readily identified by its small size and the flat and hemispherical form. Locally developed tabulate corals of *Aemulophyllum exiguum* and rugose corals of *Acrophyllum oneidaense* (Billings) have little biostratigraphic use, but the rudstone intercalations of the fragmented corals can be used as regional correlative storm beds. All the corals are endemic and cannot be correlated to European Devonian (Oliver 1976). Fagerstrom (1976) has
reported 56% of the Onondaga coral fauna occurs in the Amherstburg Formation in Ontario and Michigan. Occurrence of the pervasive chert nodules is evident in Silurian as well (Brunton and Brintnell 2011). Chert nodules often signify presence of more abundant siliceous sponge faunas and therefore elevated nutrients and associated oceanographic circulation and/or depth considerations, suggesting a slightly nutrient rich and poor circulation patterns.

Based on the occurrence of chert and these faunal assemblages, a restricted marine environment is inferred with periodic storm disturbance. A possible paleokarstic interval is suggested by the occurrence of brecciated chert nodules and solution-widened joints near the top of the Bois Blanc Formation (Fig 6.7-3).

7.2 Unnamed sandstone (below the Onondaga Formation)

At the base of Onondaga Formation, a 0–15 m thick siliciclastic sandstone unit marks an unconformity at top of the Bois Blanc Formation (Oliver 1966b). This unnamed sandstone consists of LF2, similar to the Springvale Member of the Bois Blanc Formation in lithology. Though it has long been considered the Springvale Member, the fauna of this sandstone differs from those in the Springvale Member of the Bois Blanc Formation. The fauna of fragmented rugose corals (Fig 6.7-4) (*Synaptophyllum*) and articulated brachiopods is distinctive and of Onondaga age, which helps to differentiate it from Springvale Member below the Bois Blanc Formation. In outcrop this sandstone interval is generally absent, making the Bois Blanc-Onondaga formational contact represented by a subtle lithological change that appears conformable. A greenish brown weathered, 3–5 cm thick argillaceous limestone layer may be present and marks this contact in outcrops (e.g. Dunville Quarry). Previous studies in Ontario have misidentified the lower Onondaga Formation as the top of Bois Blanc Formation (e.g. in Armstrong and Carter 2010), and drops the formational
status of this sandstone unit in Ontario. Therefore, a new name should be proposed to refer to this sandstone unit. Generally, it has a higher neutron porosity response than the overlying and underlying carbonate units thus is recognizable in geophysics logs (Fig 6.5).

Depositional History

This sandstone unit is associated with at least one erosional event prior to the cyclic deposition of the overlying Onondaga Formation. It consists of clean quartz arenite, with little argillaceous or clay minerals. Dolomite cements are absent. However, in the Springvale Member in Ontario, the dolomitic sandstone contains various clay minerals and are commonly weathered in outcrop sections.

The interpretation of its depositional environment is as reworked Oriskany sands or windblown from various provenances (the Appalachian Orogen, the northeastward Adirondacks or even the Wisconsin and/or Ozark domes to the west). The migration of the forebulge region during the non-deposition interval prior to Onondaga has removed the Lower Devonian Oriskany sandstone in the area west of Buffalo (Diecchio 1985). The Oriskany sands were then removed and redistributed into the Niagara area of the back forebulge region forming this unnamed sandstone unit on top of the Bois Blanc Formation karstic surface. The irregular thickness and the sporadic distribution of this sandstone unit is interpreted as infills of a karstic domain or reworked sand dunes/ridges on an exposed paleo-seafloor, evidenced by the sand infills in the extensive paleokarstic features at the top of the underlying Bois Blanc Formation. Fragmented fossils also indicate the rework and redistribution of carbonate skeletal grains into topographic lows by marine currents. The abundant glauconite and phosphate indicate a long time of carbonate non-production or non-deposition (hiatus in deposition) where a drowned surface was formed. Its fauna and
stratigraphic position suggests it can be correlated to the lower Amherstburg Formation in Ontario and Michigan, and possibly the Sylvania Formation in southern Ontario and Ohio.

7.3 Edgecliff Member (Onondaga Formation)

In Ontario and Western New York, the Edgecliff Member crops out along the belt from west of Buffalo to Hagarsville, and comprises light brown to light-medium grey, coarsely crystalline, massive bedded, coral-crinoidal packstone-wackestone, with bioherms locally developed (e.g. Ridgemount Quarry South; Cassa and Kissling 1982; Uyeno et al. 1982; Brett and Ver Straeten 1994; Brett et al. 2004; Armstrong 2007, 2017; Armstrong and Carter 2010). It is general 8-12 m thick, and thickens to 35 m in Core 11 OGS 82-3.

The lithofacies of the Onondaga Formation are distinctively present as allochems in finely crystalline matrix or sparry cements (Lindholm 1969). In the cored wells in Ontario, only the biostrome facies have been found in the Edgecliff Member. In outcrop sections, bioherms are present along the Onondaga Escarpment (Buehler and Tesmer 1963; Crowley and Poore1974; Cassa and Kissling 1982; Paquette 1988; Wolosz 1992). Chert nodules are more abundant in Ontario than in New York (Oliver 1956). In general, the Edgecliff is medium brown- medium to dark- grey, fossiliferous and cherty limestone with abundant corals and crinoids.

The geophysical characteristics of the Edgecliff Member yield a flat gamma-ray signature. Locally argillaceous horizons (e.g. contact of lower and middle Edgecliff members) may produce one positive gamma-ray response. The neutron log is helpful for the recognition of the Edgecliff Member by its higher values than the underlying Bois Blanc Formation.

7.3.1 Lower Edgecliff Member

The lower Edgecliff Member consists of LF3, with a relatively consistent thickness of 3–4.5
m in the Niagara region and thickening southwestward to 19 m in Core 11 (OGS 82-3) (see Fig 6.1). Its basal contact is marked either by the unnamed sandstone below the Onondaga Formation or the top of the light grey, cherty dolostone of the Bois Blanc Formation directly. When the sandstone unit is absent, the contact seems gradational and hard to place. Geophysics signatures through this contact are flat, so the contact can be only observed in cored wells or cuttings. There is a subtle change of lithology and fauna from the underlying Bois Blanc Formation into the overlying Onondaga Formation. The Bois Blanc Formation is featured by its light grey, finely crystalline, very cherty dolostone/limestone in Niagara area, with common brachiopods and rare bryozoans and corals. The lower Edgecliff Member contains abundant fossils including rugose corals, bryozoans, ostracods and trilobites that are mostly fragmented. Irregular firmground surfaces are present in the lower a few metres of this unit. Below the firmgrounds, there usually consists of 5–8 cm thick, bluish brown, bioturbated grainstone. Coral-crinoid packstone-wackestone commonly encrusts the firmground surfaces.

This unit is highly bioturbated (Fig 6.7-7), marked by vertical burrows at various levels. Storm beds consisted of fragmented rugose corals, brachiopods and crinoids in a coarsely crystalline matrix are commonly found capping this unit (Fig 6.7-8). Fauna becomes more diverse and fragmented towards the top. A good example comes from Core 2. U.S Steel No.1 from 114.7 m to 122.5 m. In this core, intact tabulate corals (Favosites) sitting on top of a finely crystalline surface are found in the lower 1 m of this unit (Fig 6.7-6), and the fossils of rugose corals, brachiopods and bryozoans also become more diverse in the upper part of unit (Fig 6.7-8). Near the top of this unit, sub-vertical microfractures and subangular breccias are evident. Fractures are exclusively filled by chert nodules. The top of this unit is marked either by a 10-cm thick, argillaceous dolostone with fractures and breccias or by a 10-15 cm thick, bluish grey, leached
rugose coral-brachiopod grainstone interval (Fig 6.8-1).

Chert nodules are common, making up 20% of this unit. The diagenetic chert nodules are irregular, replacive and displacive, tan to light grey, edged by a thin dark grey layer (Fig 6.7-6). The proportion of chert nodule is much less than that in the Bois Blanc Formation. Small, dark grey, spherical chert nodules occur within the large, light grey, irregular chert nodules, which indicate a multiple-stage growth in diagenetic process.

Although this unit has been misidentified as the top of the Bois Blanc Formation, the medium brown, coral-brachiopod wackestone unit can be distinctively differentiated from the light grey, very cherty, brachiopod mudstone-wackestone of the Bois Blanc Formation. Oliver (1966a) has claimed it is a local lithofacies of the lower Edgecliff Member in Ontario rather than Bois Blanc Formation by its distinct faunal assemblages (e.g. large Ampheginia and Synaptophyllum seen in Core 2. U.S Steel No.1) of Onondaga age.

Depositional History and Paleoecology

Similar to the Bois Blanc Formation, the lower Edgecliff lithofacies interpretation is difficult due to the presence of pervasive diagenetic chert nodules. Based on the lithological and faunal studies, this unit seems to represent several shallowing-upward cycles from above storm wave base to very shallow water within the peritidal zone or even subaerial at the top. The undisturbed tabulate corals and presence of firmground surfaces indicate a restricted shallow marine depositional environment that has been periodically influenced by sea water still stand. The breakage of rugose corals, overturning and physical erosion of corals, bryozoans and brachiopods near the top indicate a shallowing upward succession in a well circulated water above the fair-weather wave base. Minor fractures and breccias found within or plugged by chert nodules at the
top indicate a possible subaerial exposure near the end of its deposition. This subaerial duration was most likely of short duration due to lack of development of complex karstic systems.

This shallow marine deposition differs from the outer ramp lithofacies in central New York (Diecchio 1985), but is similar to the Amherstburg Formation on the Laurentian craton (Michigan Basin). Our study shows that the lower Amherstburg Formation is featured as tan to brown, slightly bituminous, very cherty wackestone to packstone with abundant rugose corals, bryozoans, brachiopods and trilobites. It is likely that this unit is correlative to the lower Amherstburg in the rest of Ontario and the sandy limestone of the uppermost Schoharie Formation in eastern New York (Vanuxem 1984; Brett et al. 1994). It is suggested that in central New York, no depositional counterpart was formed. This is possibly due to the exposure and erosion led by ephemeral uplift of the forebulge region (Ver Straeten and Brett 2000). In the subsurface of Ontario, this unit is continuously extended southward from the Onondaga outcrop belt into the Lake Erie and possibly grades westward into the lower Amherstburg Formation in the lithofacies transition zone situated between Port Burwell and Port Bruce (Fig 6.1). Cores through this zone are lacking, therefore more drilling wells and well cuttings data will be incorporated in future work to confirm the lithofacies distribution patterns.

This unit is overlain by the crinoid grainstone/packstone of LF4. The crinoid-rich interval also occurs in the middle Amherstburg Formation in Ontario (unpublished data, this study), which can represent the same later transgressive event correlative to the base of Edgecliff Member in central and eastern New York or transgressive lag deposits. Therefore, this unit is regarded as an early Amherstburg-age, restricted lagoonal deposit. This unit is older than the base of Onondaga Formation in central and eastern New York and younger than the Bois Blanc Formation in Niagara area, onlapped by the westward transgressive crinoid-dominated facies from the Appalachian
7.3.2 Middle Edgecliff Member

The middle Edgecliff consists of crinoid packstone-grainstone (LF4). This unit is 2–3.2 m thick and overlies a subaerial surface that marks the top of the lower Edgecliff Member. Large crinoid columnals and ossicles are very abundant. Chert nodules are not present. The base of this unit is very argillaceous and corresponds to a high excursion in gamma ray log (Fig 6.5). Concentrated argillaceous seams surrounding large crinoid debris marks the base of the middle Edgecliff. Rugose corals and dark grey pellets become common up-section.

Depositional History and Paleoecology

This unit represents the initial transgressive stage following the possible subaerial exposure of the lower Edgecliff. In New York, the occurrence of this crinoid grainstone/packstone usually marks the base of the Edgecliff Member (Oliver 1956). In Ontario, however, the base of Onondaga is marked by the cherty limestone facies of LF3 that overlies the unnamed sandstone below the Onondaga Formation. Previous studies show that the Onondaga sea transgressed from eastern New York where the basal Onondaga crinoid grainstone overlies the Schoharie Formation conformably (Cassa and Kissling 1982; Brett et al. 2000; Ver Straeten and Brett 2000), though in Ontario a sandstone unit marks the unconformity between the Onondaga Formation and the underlying Bois Blanc Formation. The crinoid-rich bed can be used as a regionally correlative marker bed because the crinoid ossicles tend to fall apart on death and then be distributed by water waves or currents to form a regionally extensive bed. This crinoid packstone/grainstone facies transgressed the lower Edgecliff in Ontario, suggesting the seaway has connected the Appalachian Basin to the Laurentia craton interior in Niagara area over the forebulge region to the east.
7.3.3 Upper Edgecliff Member

The upper Edgecliff Member is present as biostromal lithofacies (LF5). Its basal contact is marked by a lithological change from the underlying crinoid packstone to very fossiliferous cherty wackestone-packstone. An abundance of rugose corals of *Acinophyllum* and *Synaptophyllum* marks the base of this unit. The thickness of the biostromal facies ranges from 1.5 m to 3 m.

The biostromal facies (LF5) recognized in cored wells consists of abundant fossils, including a diverse fauna of predominant rugose corals (*Acinophyllum, small Heterophrentis, Cystiphyllum, Synaptophyllum* and *Cylindrophyllum*) with common tabulate corals (*Favosites* and rare *Syringopora* and *Cladopora*) (Fig 6.8-3). Other fossils, including crinoids, bryozoans and *Thamnophora* are rare. Lenticular stromatoporoids are present at the top of this unit (Fig 6.8-4). Matrix is fine-grained or sparry cement, and the fossils are thought to have formed *in situ* or not transported far. Intercrystalline porosity in the tabulate corallites partly plugged by lime-muds and sparry cements, forming geopetal fabrics. Large *Heterophrentis* has been found at Morgan Point in Ontario (Fig 7.1-3). The intercrystalline porosity within corals may form an interconnected reservoir system by the evidence of large amount of oil stains having been found in outcrops in Norfolk Quarry (7.1-6) and in Core 5 (Consumers’ Amoco 130102) (Fig 6.8-5).

Concentrated argillaceous stylo-seams are commonly found surrounding the chert nodules and fossil edges. They are interpreted as diagenetic pressure-dissolution residues. When the displacive chert nodules grow, they create great pressure to dissolve the limestone matrix especially along the fossils edges, then the undissolved clay residues are left to form the concentrated and wispy stylo-seams (Folk and Weaver 1952).

The upper contact with the mudstone in Clarence Member is marked by the first occurrence of dark grey mottled chert nodules (Fig 6.8-6). The chert nodules are composed of cryptocrystalline quartzs with a higher organic content (Parkins 1977) observed in thin sections. In some cores, the
cherty mudstone in the Clarence Member interfingers with the fossiliferous wackestone-packstone of the upper Edgecliff Member (e.g. in Core 2 U.S. Steel No.1), which indicates the Clarence Member should be a local lithofacies at the top of the Edgecliff Member.

The bioherm facies crops out in the Hagarsville area and in the Long Point area in Ontario as indicated by Wolosz (1991), though it has not been recognized in cored wells in our database. A well-studied example comes from the LeRoy area west of Buffalo in western New York (Crowley and Poore 1974; Wolosz 1991, 1995a). Small bioherms, 6 m thick and 30 m wide, have been observed in Ridgemount Quarry South (Brett et al. 2004) (Fig 7.5). Most Devonian reefs are formed by stromatoporoids and tabulate corals (Copper 2002). In Ontario, however, the Edgecliff bioherms are formed by colonial rugose corals (*Acinophyllum*), small tabulate corals (*Cladopora* and *Thamnopora*) and crinoids (Crowley and Poore 1974), and no stromatoporoids have been observed as bioherm frame builders. Brachiopods, molluscs, sponges and unrecognizable fossil fragments have also been found within the bioherm facies in western New York (Crowley and Poore 1974; Lindermann 1988; Friedman 1995; Bruner and Smosna 2002). Wolosz (1992, 1995a) have described two paleocommunities of the Edgecliff bioherms. The first contains phaceloid rugosan mounds and thickets, and the second consists of favositid corals and crinoids bank. The rugosan mounds and thickets are usually 6–15 m thick, up to 200 m in diameter (Wolosz 1991) and the fauna diversity within is low. The favositid coral-crinoid bank is dominated by favositid corals, though they have not developed a framework mass to support a reefal structure. Crinoid debris are very common among the coral communities. Solitary rugose corals and bryozoans are also present, and delicate branching corals of *Cladopora* and *Syringopora* are present (Lindermann 1988).

In western New York, the biohermal facies are different in fauna from those in central and
eastern New York. It is dominated by tabulate corals of Cladopora and Thamnopora, with a base formed by colonial rugose corals of Acinophyllum. The best studied LeRoy bioherm facies consists of 6 m thick colonial rugose corals and crinoids, and is 150 m in diameter (Lindermann 1988). The bioherm in Ridgemount Quarry South is 5.5 m thick, and the lateral extent is unknown (Fig 7.5). Four biohermal facies are observed in Ridgemount Quarry South: basal facies, core facies, flank facies and caprock facies. The bioherm began with a base of colonization of Acinophyllum corals. The infill of disaggregated lime muds into the corallite framework comprises the basal facies of the bioherm. Core facies comprises massive, medium grey bafflestone of predominant Cladopora. Other fossil debris and lime muds comprise the matrix and an encrusting fenestral bryozoan (Fistulipora?) is present (Brett et al. 2004). Flank facies on the western side (leewards) is formed by alternate layers of crinoidal packstones-wackestone and fossiliferous rudstone of Acinophyllum, Cladopora, Heliophyllum and other rugose corals. The cap of the bioherm is apparent on top of the Cladopora-rich core facies. It comprises colonial rugose corals and lenticular stromatoporoids. Other fossil debris may be locally common. The skeletal grain size is much smaller than the core facies. Platycerid gastropods are often present in the overlying unit of the bioherm (i.e. in Nedrow/Clarence Member). The occurrence of platycerid gastropods suggests an impoverished fauna of harsh environment (Oliver 1956). As predators or parasites of corals and crinoids, the gastropods are probably a commensalistic partner of the other biohermal fauna.

Depositional History and Paleoecology

In the upper Edgecliff Member, the biostromal/biohermal indicates that the transgression having begun at the base of the middle Edgecliff Member may have continued into the late Edgecliff-time (Koch 1982). The biostromes were likely to have formed in between the major
bioherm build-ups. During this time, eastern and western New York areas were covered by shallow waters where coral-rich facies thrived (Oliver 1954). The locally developed bioherms have created an irregular surface on the ramp in Ontario and western New York. The Niagara area was interpreted to be situated on the inner-middle ramp, and the fauna in the biostromal/biohermal facies in Ontario represents a shallower and warmer water realm than that in central and eastern New York.

Interpretations of the paleobathymetry of the bioherms/biostromes vary among previous researchers, because unlike the other Devonian buildups, the framework of the Onondaga bioherms are not built by stromatoporoids/tabulate corals and encrusting algae (Cassa and Kissling 1982). No subaerial surfaces or hummocky beds have been recognized either. Wolosz (1995a) suggested that the rugosan mounds/thickets were formed in relatively shallow waters with high turbidity, and the coral-crinoid bank was formed in offshore deeper waters where the turbidity is not high enough to support the growth of rugose coral communities. However, the rugosan mound/thickets and the coral-crinoid bank are not mutually exclusive. He suggested that coral-crinoid bank facies atop the rugosan mounds marked a transition from shallow to deep water depositional environment. This seems unlikely because no correlative deepening cycle has been found in Ontario, and the occurrence of tabulate corals and stromatoporoids on top of the biostromal facies (e.g. Core 10. Consumers’ Pan Am 13057) indicate a shallowing-upward pattern rather than deepening-upward. Ver Straeten and Brett (2000) pointed out that the westward migration of the short-lived forebulge has controlled the distribution of the Edgecliff bioherms. The bioherms tend to build upon topographic highs of the forebulge region and extend only a few hundred metres in diameter. Therefore, the bioherms are thought as shallow water, isolated build-ups. Koch (1981, 1982) and Koch and Boucot (1982), based upon brachiopod assemblages, have
inferred that the bioherms and mounds grew below the photic zone. Cassa and Kissling (1982) have raised up an alternative interpretation of very deep water bioherm model (>200 m deep) by comparing the Edgecliff bioherms with the modern ahermatypic corals in Norway in the Arctic ocean (Teichert 1958). This is also unlikely because the paleolatitude of the study area was 30-35° south of the paleoequator (Fig 1.6) and the fauna of tabulate corals and stromatoporoids in Ontario would never grow in such cold and deep water below the carbonate compensation depth (CCD) (Copper 1994).

Wolosz (1991) also suggested a cool, subtropical environment for the growth of the Edgecliff bioherms based upon the lack of calcified algae and sponges (stromatoporoids), pisoids, ooids and hummocky cross-bedding in New York. In Ontario, the biostromal/biohermal facies are thought to form in shallower water at a tropical (Fig 1.6), up-ramp environment in an epicratonic seaway, because of the occurrence of large amount of tabulate corals and a small portion of stromatorporoids. It is likely that the Appalachian Basin was never too deep during Onondaga-time and the distribution of fauna assemblages were controlled by the paleolatitude and forebulge migration.

7.4 Clarence/Nedrow Member (Onondaga Formation)

In the type area in central New York, the Nedrow Member overlies the Edgecliff Member conformably and comprises an argillaceous, cherty limestone with shale interbeds and a sparse fauna of platycerid gastropods (Oliver 1954, 1955; Cassa and Kissling 1982; Brett and Ver Straeten 1994). In eastern and central New York, the Nedrow Member represents the deepest water deposits within Onondaga Formation. The lower Nedrow consists of platycerid gastropods and brachiopods fauna, and goniatites have been found (Oliver 1956). Near the top, a pair of black shale interbeds
marks the top of the Nedrow Member (Brett and Ver Straeten 1994), underlying the fossiliferous Moorehouse Member.

In western New York and Ontario, the unit overlying the Edgecliff Member is faunally and lithologically different from the Nedrow Member and therefore has been given a different name (Clarence Member) (Oliver 1966a; Mesolella 1978; Rickard 1981). It is featured as olive grey, sparsely fossiliferous argillaceous carbonate mudstone with abundant dark chert nodules (LF6). It is well exposed at Thunder Bay east in Ontario (Fig 7.3) with weathered dark grey chert nodules and trace fossils of Zoophycos. Recent studies show that the Clarence Member cherty, micritic facies grade laterally eastward into the middle and upper Edgecliff and interfinger with upper Edgecliff and lower Moorehouse members (Brett and Ver Straeten 1994; Brett et al. 2011). In this study, the Clarence Member is retained for the very cherty and sparsely fossiliferous limestone facies that separates the Edgecliff from the overlying Moorehouse members.

Its finely crystalline matrix and sparse fauna of gastropods, rugose corals, brachiopods and bryozoans indicate the deepest water deposits in the Onondaga Formation. The abundant diagenetic black chert nodules, that make up more than 50% of the Clarence Member. In thin sections, these dark grey chert nodules reveal higher organic content than the light grey chert nodules in the Bois Blanc Formation, suggesting a low sedimentation rate for the Clarence Member. The thickness of Clarence Member varies from 8-14 m in western New York to less than 5 m in the subsurface of Ontario. It is present everywhere overlying biostromal lithofacies of the Edgecliff Member, but in the biohermal outcrops the Edgecliff Member is overlain by the base of the Moorehouse Member, and the cherty mudstone of Clarence is absent (Crowley and Poore 1974).

In the subsurface of Ontario, the Clarence Member is present as interbeds with the upper Edgecliff and lower Moorehouse members, and the contact appears to be gradational (e.g. in core
Consumers’ Pan Am 13058, Fig 6.8-6). A pair of chert-rich mudstone interval interbeds with the upper Edgecliff biostromal facies have been found in almost all collected cored wells. In the subsurface of Ontario, the Clarence Member is more fossiliferous than that in western New York. Fragments of brachiopods and trilobites are sparsely present and the mudstone matrix is highly bioturbated (Fig 6.6-LF6) by gastropods. Geophysics characteristics of Clarence/Nedrow Member is distinct in its type area with a high kick of gamma-ray signature (the highest within Onondaga) and a decrease in neutron porosity because of the presence of shale layers. In the subsurface of Ontario, the Clarence Member is cleaner (less terrigenous mineral components) and displays a flat gamma-ray response, but it still displays a decrease in neutron log that can be recognized regionally.

Depositional History and Paleoeology

Interpretation of the depositional environment of Nedrow/Clarence Member varies. Koch (1982) thought the Nedrow Member represents a regression following the deposition of Edgecliff because it forms a sharp basal contact with the fossiliferous Edgecliff Member with a severe increase in argillaceous materials that was harsh for the growth of corals and crinoids. A regression model also matches the shallowing-upward cycle from the lower restricted lagoonal Amherstburg Formation grading upward into the evaporitic Lucas Formation in the Michigan (Koch 1982; Uyeno et al. 1982). Therefore, shale top of the Nedrow Member may not have been formed in deeper-water environments. Rather, it could have been the result of the lowering of sea level that led to the sea water stagnation and a low carbonate production rate (see Gurney and Friedman 1986), and the decrease of faunal diversity and the disappearance of corals and crinoids are due to the influx of clay and mud during the regression (Koch 1982). However, this hypothesis is largely based on the correlation with the peritidal-sabkha Lucas Formation.
Brett and Ver Staenten (1994) have suggested that Nedrow represents a sea level rise and the presence of shale at the top of Nedrow Member marks the maximum flooding surface. They traced the shale in central New York to east of Buffalo, but could not trace it any further westward. This supports the idea that strata in Ontario may represent an up-ramp position during the deposition of Onondaga Formation. It is likely that the shales pinch out in western New York and Ontario because of a westward shallowing depositional environment. Evidence also comes from the higher fossil content in Clarence Member in Ontario than in New York (Fig 6.8-6).

The Nedrow Member is dominated by platycerid gastropods (Oliver 1954), which is a faunal group that can tolerate high influx of siliciclastic materials. Tentaculites and goniatites have been reported by Oliver (1956) and Lindemann and Yochelson (1984). The lack of corals and crinoids may be a result of drowning event during the sea level rise. In Ontario, the Clarence Member possesses more brachiopod and trilobite fossils, and the color of chert nodules becomes lighter southwestwards. This indicates a decrease of organic matter burial in Clarence Member and a shallower water environment located at up-ramp. Therefore, the Clarence/Nedrow Member represents a deepening-upward succession overlying the Edgecliff Member.

7.5 Moorehouse Member (Onondaga Formation)

The Moorehouse Member was named by Oliver (1954) in outcrop section 2 km southeast of Syracuse, New York. In this type section, it comprises dark grey, cherty, finely crystalline limestones with calcareous shale beds that represent an outer ramp setting. Its basal contact is marked by the occurrence of fossiliferous limestone overlying the two shale beds at the top of the underlying Nedrow Member (Brett and Ver Straeten 1994). These shales can be traced northeastward into Pennsylvania (Ver Straeten 2007). Potassium-rich ash beds occur at several
levels within the upper Moorehouse Member, and the Tioga ash bed marks the upper contact with
the Seneca Member (Dennison and Textoris 1977; Conkin and Conkin 1979; Roen 1980).

In the subsurface of Ontario, however, only the lower Moorehouse Member is present and
comprises LF7 and LF8. It represents a return to inner-middle ramp deposition; shale layers are
absent in the lower Moorehouse Member. Although the fossiliferous facies of LF7 resembles the
biostromal facies of upper Edgecliff Member (LF5), the lack of colonial rugose corals
(Acinophyllum) and the occurrence of laminar and bulbous stromatoporoids differentiate it from
the LF5. Fossils are dominated by solitary rugose corals (Eridophyllum and Heliophyllum),
brachiopods and rare tabulate corals. No bioherm or reefal facies have been reported in
Moorehouse Member (Wolosz 1991). LF7 is less argillaceous but chert is much more common
than the unit in central and eastern New York. A crinoid-brachiopod grainstone facies (LF8) marks
the top of the lower Moorehouse Member (Fig 6.5-LF8). LF8 represents the shallowest deposition
in Moorehouse Member that usually caps each shallowing upward cycle and represents the
termination of Moorehouse Member deposition in Ontario. The crinoids and brachiopods have
been possibly transported by waves or storms and formed shelly-crinoidal shoals in middle
Moorehouse Member. Leaching and dolomitization prevail up-section, and porosity decreases due
to cementation by sparry calcite/dolomites. An irregular unconformity separates it from the
overlying Dundee Formation. In New York, this unconformity is recognizable regionally and is
overlain by the dark grey fossiliferous limestone of upper Moorehouse (Oliver 1956; Brett et al.
2011). The upper contact with the overlying Dundee Formation is a distinctive unconformity that
is overlain by a 0.5 to 2.5 m thick sandy limestone or sandstone interval at the base of the Dundee
Formation (e.g. in Core 5. Consumers’ Amoco 13102, Fig 6.10-2) or a 10-cm thick argillaceous
mudstone layer that has been interpreted as an undissolved paleokarstic surface (e.g. in Core 4.
Consumers’ Amoco 13061, Fig 6.6-LF8).

The upper Moorehouse Member contains up to four shale layers in its type section, and is called “False Nedrow Shale” by previous workers (Brett and Ver Straeten 1994). No shale layers/pulses in the Moorehouse Member have been reported in Ontario. The Moorehouse Member has a flat response of gamma-ray log and neutron porosity. At the upper contact, the gamma-ray radiation remains low but neutron porosity decreases into the Dundee Formation.

Depositional History and Paleoecology

The lower Moorehouse Member in Ontario represents shallower water deposition than its deeper basinal counterpart in central New York, with a pronounced break at the Moorehouse Member – Dundee Formation contact that can be correlated to the unconformity in the mid-Moorehouse in central and eastern New York (Oliver 1956; Mesolella 1978). Koch (1981; 1982) interpreted the lower Moorehouse as a transgression terminated by the Tioga ash bed, while Gurney and Friedman (1986) and Brett et al. (2011) thought the lower Moorehouse represents a regression stage, based on the increase in coarser grained crinoid-brachiopod skeletal materials in finely crystalline matrix and caps by a sequence boundary of mid-Moorehouse unconformity. The upper Moorehouse represents a transgression marked by influx of fine-grained carbonate materials of bryozoans and brachiopod fragments (Brett et al. 2011). In Ontario, a shallow marine Atrypid-Megakoziowkiella brachiopod community in the lower Moorehouse has been reported by Boucot (1975), and in the central and eastern New York the lower Moorehouse consists of a slope community Atrypid-Levenea. Therefore, the lower Moorehouse in Ontario indicates a typical inner-middle ramp environment deposits. A regression with minor fluctuation of sea level changes occurs through to the end of the lower Moorehouse deposition. The LF8 represents the shallowest
deposits of shoal facies, which occurs at two levels near the top that mark the termination of carbonate production. The faunal diversity decreases towards the top, which indicates a sea level drop and the end of lower Moorehouse deposition.

7.6 Seneca Member (Onondaga Formation)

The Seneca Member is only present in central and eastern New York (Oliver 1954; 1956; Wolosz and Paquette 1988). Its lithology is slightly different from that of the Moorehouse Member (Brett and Ver Straeten 1994), in that it is lighter grey, more argillaceous and the skeletal fragments are smaller (Mesolella 1978; Rickard 1981). No chert nodules are present (Oliver 1956). Brachiopods and crinoids are dominant in the Seneca Member, and crinoid content decreases upward (Feldman 1985). Fish bone beds have been reported by Martin (2002) who interpreted these beds as transgressive lag deposits indicating a deepening cycle during the Seneca-time. He also correlated them to the bone beds in Delaware Limestone in northern Ohio (see Leonard 1996). Birchard (1993) has reported a fish bone bed in the Dundee Formation in Ontario that has also occurred in the Seneca Member in New York. The Seneca Member is overlain by the shales of Marcellus Formation by a sharp contact (Feldman 1980), which has a positive trend in gamma ray log signature.

Depositional History

Koch (1981) interpreted the Seneca Member as a regressive unit following the Moorehouse transgressive succession. Brett et al. (2011) and Feldmann (1980) have interpreted the upper Moorehouse and Seneca members to represent a deepening upward cycle into the Marcellus shale. The faunal diversity decreases upward and the presence of deep water deposits of Marcellus shale
may support a deepening-upward hypothesis of the Seneca deposition in New York.

7.7 Dundee Formation

No upper Moorehouse and Seneca members lithofacies are present in Ontario. The Dundee Formation in Niagara area is thought as equivalent to the upper Moorehouse and Seneca in Ontario (Sparling 1985). Oliver (1956) found a lithological transition from light grey, non-cherty Seneca limestone to the light brown, cherty, crinoidal limestone of Dundee Formation west of Buffalo. It is thus inferred that the Dundee Formation may be an equivalent unit of the upper Moorehouse and Seneca members in Ontario and Michigan. Conodont studies by Oliver and Klapper (1982) and Uyeno et al. (1982) also support this correlation. Though not observed in the cored wells, the Tioga ash bed can be recognized in gamma-ray logs by a distinct narrow positive excursion in the Dundee Formation in Niagara area. It is unfortunate that none of the cored wells have captured the top of the Dundee Formation in the Niagara area where the Tioga ash bed may be present. However, a 3-cm thick, biotite-rich ash bed has been discovered in Core 5 (Consumers’ Amoco 13102) at the depth of 107.5 5m (Fig 6.10-6), which is 24 m above the base of Dundee Formation. In the adjacent cored wells of Core 11 (OGS 82-3) (Fig 6.10-7) and Core 9 (Consumers’ Amoco 13076) (Fig 6.10-8) that are 20 and 35 km southwest of the Onondaga deposition edge (Fig 6.1) separately, the Tioga ash bed has been recognized about 20 m above the base of the Dundee Formation. This ash bed in Ontario has similar mineralogy components with that of the Tioga ash bed in New York, Indiana and Pennsylvania. More dating test is needed for further correlation.

Droste and Vitaliano (1973) first reported the Middle Devonian Tioga K-Bentonite bed in Indiana. Baltrusaitis (1974) was the first to find the Kawkawlin bentonite bed in Michigan Basin and he suggested that it is correlative with the Tioga Bentonite in Indiana. Dennison and Textoris
(1978) discovered the Tioga K-Bentonite time-marker in the Onondaga Formation in the Appalachian Basin. In eastern Ohio, Collins (1979) discovered the same bentonite bed that marked the contact of Columbus and Delaware formations. Discussions on the formation of this time-marker bed was focused on its mineral composition. Commonly a bentonite is consisted of dominantly illite and kaolinite. However, Roen and Hosternman (1982) challenged the usage of “bentonite” referred to this ash bed by the fact that it is featured as dark grey, biotite-rich unit with rare illite minerals and no kaolinite. Therefore, the name “Ash Bed” is more appropriate to be applied to this marker bed instead of “Bentonite”.

Based on these discoveries, Roen (1980) correlated the Tioga ash bed in New York into the Michigan Basin, Indiana and Ohio and suggested that they were formed by the same volcanic eruption in Indiana. Overall, the discovery of the biotite-rich ash bed in upper Dundee Formation in Niagara area has provided evidence for its correlation to the upper Moorehouse and Seneca members in New York and the upper Columbus and Delaware formations in Ohio.

Lithology

In Ontario, the Dundee Formation is a 35-42 m thick limestone unit between the overlying shales of the Hamilton Group or Marcellus Formation and the underlying limestone/dolostone of the Lucas Formation (Sanford 1967; Johnson et al. 1992; Uyeno et al. 1982). The primary depositional facies of the Dundee Formation have been described in detail by Birchard (1990, 1993). It is featured as medium grey to brown, fossiliferous, crinoid-rich limestone with common chert nodules (Johnson et al. 1992). Its basal contact with the Lucas Formation is marked by an irregular unconformity (Armstrong and Carter 2010). Siliciclastic sand grains are present in Huron and Oxford counties at the base of the Dundee Formation (Birchard 1990). Six lithofacies have
been recognized in SW Ontario (Birchard 1990, 1993), and the general depositional pattern shows a deepening upward cycle that grades into the brachiopod-rich shale of Marcellus Formation (Rickard 1984; Johnson et al. 1992). Although regional studies have been undertaken well by Birchard (1990, 1993), no cored wells of the Dundee Formation overlying the Onondaga Formation in Niagara area were incorporated in that study. In this study, the identified lithofacies are chiefly based on Birchard’s (1990; 1993) classification schemes, with emphasis on various fauna assemblages, primary sedimentary structures and occurrence levels of ash beds.

7.7.1 Lower Dundee Formation

The lower Dundee Formation consists of LF9 to LF11 in Niagara area. The basal contact is marked by an irregular unconformity that separates the lower dark grey, fossiliferous limestone and the overlying sandy, cherty, light brown, crinoid-rich, finely crystalline limestone of Dundee Formation (Fig 6.10-1 to 6.10-2). Locally, sand grains may be found sparsely dispersed in the limestone matrix, at the base of the Dundee Formation (Fig 6.10-2). A 3-4.5 m thick unit of LF9 usually overlies the sandy base and grades into the LF10 up-section. LF9 is highly bioturbated by the presence of oval to irregular bioturbation marks (Fig 6.10-3). It is slightly argillaceous, and wispy stylo-seams are locally concentrated around the diagenetic chert nodules. This stylo-seam bounded nature has given this lithofacies a pseudonodular texture that has been reported by Birchard (1993). The light grey, parallel-bedding or irregular chert nodules make up less than 10% of LF9. Fossils include very common crinoids, brachiopods and small solitary rugose corals. LF9 is usually overlain by a 20–25 cm thick, cream to light grey, fenestral-pelloital grainstone interval with very common fossil fragments (LF10). In this lithofacies, the winnowed fossil fragments of brachiopods, corals, crinoids and gastropods are well cemented by sparry calcite (Fig 6.9-7). Peloids are the predominant skeletal grains in this lithofacies. Porosity is very high, possibly
enhanced by dissolution. Interparticle and intraparticle porosity has been observed in thin section that makes up around 40% volume of the overall rock. LF10 is commonly overlain by the crinoid-coral wackestone-packestone of LF11 (Fig 6.9-7, 6.9-8 and 6.10-1). Crinoids are predominant in this lithofacies. Packstone pulses are consisted of locally concentrated crinoid debris. Tabulate corals (*Favosites*) and small rugose corals are present in multiple levels. Irregular, white chert nodules make up less than 10% of this lithofacies. LF11 grades into the dark brown, bituminous mudstone with rare fossils of LF12. Interbeds of LF11 and LF12 could be recognized at the top of the lower Dundee Formation.

*Depositional History*

The lower Dundee Formation represents a deepening-upward succession. The LF9 and LF10 are interpreted as protected shallow marine environment with periodic shallow peritidal deposits. The LF11 indicates an open marginal marine, middle-ramp condition.

The basal transgression surface is marked by the eolian siliciclastic sand grains dispersed in the *in situ* limestone matrix. Both peloids and bioturbation indicate a protected shallow marine setting. Periodic peritidal depositional environments are suggested by the occurrence of LF10. Winnowed fossil fragments and common peloids indicate a shallow marine subtidal to intertidal environment with dissolution enhanced porosity. Deepening-upward cycle starts with the occurrence of LF11. Crinoids and corals represent a shallow open marine environment with no depositional break or evidence of subaerial exposure. Corals are rare in the Dundee Formation in southern Ontario and Michigan, but they are commonly present in Niagara area and western New York. The lower Dundee Formation in Niagara area is also less bituminous and more argillaceous. This may suggest a middle ramp depositional environment that deepens into the Appalachian Basin.
axis eastward and grades into the upper Moorehouse Member.

**7.7.2 Middle Dundee Formation**

The middle Dundee Formation consists of dominantly LF12, interbedded with LF11 and LF13. Such stratigraphic relationships can be compared to the lithofacies 3 described by Birchard (1993). The LF12 is thinly bedded to massive mudstone with rare black chert nodules (Fig 6.10-6). It overlies the LF11 by a sharp contact. Individual mudstone beds are 15-50 cm thick. Dark grey, carbonaceous seams are common. Megafossils are rare. Fragments of rugose corals, brachiopod and crinoids are found, and *Tentaculites* and spore case (*Tasmanites*) are present. Chertified gastropods has been found in Core 7 (BH08-21). The top of this unit is marked by the light grey, very fossiliferous limestone of LF13. Content of chert nodules decreases upwards, and in LF13 no chert nodules are present.

*Depositional History*

The LF12 represents restricted lagoonal, quiet water deposits indicated by the rare megafossil contents and the occurrence of *Tasmanites* (Dutta et al. 2006). The thinly bedded, dark brown mudstone suggests a quiet water, possibly dysaerobic environment. The dark carbonaceous seams and dark chert nodules indicate a burial of large amount of organic carbon. The absence of marine suspension feeders and filter feeders indicates stressful benthic conditions. Fragmented megafossils may have been transported into the depositional environment by storms.

**7.7.3 Upper Dundee Formation**

Upper Dundee Formation comprises of entirely LF13. In Niagara area, the top of the Dundee Formation is not cored, so the upper contact with the overlying units are not observable. The
lithology of the upper Dundee Formation generally comprises brachiopod-crinoid wackestone to packstone (Fig 6.10-7 and 6.10-8). Pyritization and possible firmgrounds have been described by Birchard (1990, 1993). Megafossils are dominated by crinoids and brachiopods, with rare trilobites and rugose corals. Fossil content and diversity decreases upwards. Grainstone intervals are locally present, representing minor sea level fluctuation. Chert nodules are very rare. Possible Tioga ash beds are recognized the upper Dundee Formation (Fig 6.9-5 to 7).

Depositional History

This unit represents a deeper water, open outer ramp setting in contrast with the overlying restricted lagoon facies. Periodic reworking of skeletal materials produced the fossiliferous grainstone facies within LF13. In general, it records a deepening-upward cycle, as suggested by its darker color and decreased fossil diversity. Overlying the ash bed, fossils become less common and the limestone matrix is finer crystalline. This deepening pattern corresponds to the lithological change from upper Moorehouse Member and the Seneca Member in central New York.
Figure 7.1 Lithology of the Edgecliff Member of the Onondaga Formation in outcrops.

1. Photograph showing large crinoid columnals and solitary rugose corals (*Heterophrentis*), a typical fauna assemblage of basal strata of the Edgecliff Member in New York and Ontario in outcrops. Hammer scale is 15 cm in width.  
*(Ridgemount Quarry South)*

2. Biostromal facies of the upper Edgecliff Member. Colonial rugose corals (*Acinophyllum*) are abundant in the dark grey, finely crystalline limestone matrix. Bioherm facies is not present and large chert nodules are very common. This facies usually forms the base of the bioherm build-ups (*see* Figure 7.5). Hammer scale is 33 cm in length. *(Ridgemount Quarry South)*
**Figure 7.2** Contact of the Onondaga Formation and the Bois Blanc Formation and the lithology of Edgecliff Member in outcrops.

1. Contact of the lower Bois Blanc Formation and the upper Onondaga Formation. The unconformable contact is marked by a sharp clayey surface. Both units are featured as light grey cherty limestone. A more diverse fauna is present in the overlying Edgecliff Member. ([Ridgemount Quarry South](#))

2. Large colonies of the rugose coral *Acinophyllum* in the Edgecliff Member. *Acinophyllum* is the predominant component in bioherm build-ups. ([Morgan’s Point](#))
Figure 7.3 Lithology of the Clarence Member and the Moorehouse Member in Ontario.

1. Weathered surface of the Clarence Member. The dark grey chert nodules are less susceptible to dissolution thus formed an irregular surface. Trace fossils of *Zoophycos* are present. (Thunder Bay East)

2. Weathered surface of the Moorehouse Member. Lithology resembles the upper Edgecliff, but the fauna is less diverse. Another difference is that colonial rugose corals are absent. (Selkirk-Hemlock Creek)
**Figure 7.4** Contact of the Moorehouse Member of the Onondaga Formation and the Dundee Formation in outcrops, Ontario.

1. The cherty, fossiliferous Moorehouse Member is overlain by the buff to light grey, cherty limestone of the Dundee Formation. The unconformable contact is flat and sharp. The basal 25 cm of the Dundee Formation is sandy and less cherty than the underlying Moorehouse Member. Note the cycles of medium bedded, fossiliferous wackestone and the thinly bedded argillaceous mudstone in the uppermost Moorehouse Member. *(Norfolk Quarry)*

2. Contact of the Moorehouse Member and the Dundee Formation as exposed in the eastern wall of the Norfolk Quarry. The upper contact surface is overlain by a 5-cm thick argillaceous layer. Iron stains indicate a possible water flow zone formed along this unconformity. *(Selkirk-Hemlock Creek)*
**Figure 7.5** Biohermal and mound facies of the Edgecliff Member in Ridgemount Quarry South

1. Bioherm facies in the Edgecliff Member described by Brett et al. (2000). This bioherm is 15 m in width and 6 m in height. Bioherm facies, including base facies, core facies, flank facies and cap facies are present. *(Ridgemount Quarry South)*

2. A light grey, fossiliferous mound is present in the upper Edgecliff Member that is overlain by the dark grey, argillaceous limestone. *(Ridgemount Quarry South)*
7.8 Subsurface Correlations

Based on the stratigraphic descriptions and interpretations above, it can be concluded that the Onondaga Formation in the Niagara area can be correlated with the lower three members of the Onondaga Formation in New York. As such, the nomenclature of the “Onondaga Formation” should be established to refer to this dark grey, fossiliferous limestone unit in Niagara area. The name Amherstburg Formation should not be applied in the Niagara area because it has a different lithology and represents a different depositional environment that of the Detroit River Group. In addition, the Onondaga Formation in Ontario is not entirely correlative to the Amherstburg Formation.

The correlation of the Onondaga Formation into the Michigan Basin has long been problematic (e.g. Sanford 1967; Rickard 1975, 1984; Sparling 1985; Brett et al. 2011). In the subsurface well data of SW Ontario, the major lithofacies changes from Onondaga Formation to Detroit River Group occur in the area that covers Core 5 (Amoco 13102) to Core 10 (Pan Am 13057) and Core 11 (Amoco OGS 82-3), extending from Port Burwell westward to Port Bruce (Fig 6.1). West of this transitional zone, the Onondaga Formation is not recognizable, but the Amherstburg Formation and the Anderdon Member of the Lucas Formation are present (Fig 7.6).

In core 5 all three members of the Onondaga Formation are present, and the Moorehouse Member contains a stromatoporoid-\textit{Amphipora} (calcified sponge) fauna that is different from that of the Moorehouse Member in western New York. The amphiporid mounds are also present in the Lucas Formation in Amherstburg Quarry (Birchard et al. 2004), which may represent a regionally correlative bed from the Lucas Formation to the Moorehouse Member.

In Core 10, all three members of the Onondaga Formation and the Anderdon Member of the Lucas Formation are present. The crinoidal grainstone at the top of Moorehouse Member is
overlain by tan to light brown, medium to thickly bedded lime mudstone to wackestone of the Anderdon Member (Lucas Formation) that contains common fragments of bulbous stromatoporoids, which represents a lagoonal or back-reef facies of the fossiliferous Moorehouse lithofacies in western New York. Other evidence to support the correlation of the Anderdon Member of the Lucas Formation to the Moorehouse Member of the Onondaga Formation comes from conodont studies (Telford et al. 1977; Oliver and Klapper 1981; Uyeno et al. 1982) showing that both belong to the same lower *costatus costatus* conodont biozone.

In Core 11 that is 20 km west of core 10, the Onondaga Formation is not recognizable and the lithologies have been identified as Detroit River Group rather than Onondaga Formation. The units present in this core between the overlying Dundee Formation and the underlying Bois Blanc are, in ascending order, a sandstone unit below the Amherstburg Formation; the lower Amherstburg (cherty coral-crinoidal wackestone); the upper Amherstburg (biostromal stromatoporoid-*Amphipora*-coral packstone-wackestone); and the Anderdon Member of the Lucas Formation (interbeds of stromatoporoid-coral wackestone and stromatoporoid-coral floatstone). Though referred to as Amherstburg Formation, the upper Amherstburg biostromal facies contains similar fauna of stromatoporoids and *Amphipora* to that in the Moorehouse and Edgecliff Members in Core 10. The Clarence Member is not present in Core 11 and the lower Amherstburg Formation comprises cherty wackestone that resembles the lower Edgeliff Member in Niagara area and the lower Amherstburg Formation from subsurface southwest of the Algonquin Arch and in Michigan, both lithologically and faunally. Such a lithofacies transition may indicate that the lower Amherstburg Formation represents a regionally distributed intracratonic deposit, and the biostromal facies or reef facies at the top of the Amherstburg Formation (including the Formosa Reef) are possibly correlative to the upper Edgecliff or lower Moorehouse Member of
the Onondaga Formation because of the faunal similarity (lenticular and bulbous stromatoporoids, rugose and tabulate corals and brachiopods).

The fragments of stromatoporoids and corals in the Anderdon Member of the Lucas Formation in Core 11, which overlies the biostromal/reefal facies of Amherstburg Formation, were possibly shed from the reef tops or the biostromes on a paleo-topographic high to the east, or from multiple phases of mounds that backstepped on the ramp during relative sea level rise. In Core 10, the Anderdon Member also comprises cycles of lagoonal facies interbedded with the Moorehouse Member biostromal facies that were controlled by local sea level changes and possibly paleotopographic variations. Based on the lateral lithofacies change pattern within these three cores, it is likely that the Detroit River Group represents shallower water, cratonic interior deposits that can be correlated with the Onondaga Formation within a third-level sequence. It is inferred that the depositional environment became more restricted towards the west into the Michigan Basin, evidenced by the occurrence of evaporites and dolomite interbeds in the Lucas Formation west and north of the St. Mary’s and London area into the craton interior in Michigan (Birchard et al. 2004). The coupling of the forebulge migration and biohermal growth in western New York (Ver Straeten and Brett 2000) probably played a role as barriers to form a lagoonal depositional realm westward into Ontario where the stromatoporoid-Amphipora framestone in the uppermost Amherstburg Formation and more sparsely fossiliferous Lucas Formation (e.g., Anderdon Member) are evident. Problems remain, however, in this correlation, including the correlation of various stratigraphic levels of siliciclastic sand units across the forebulge region.

Another challenge is whether the open marine facies of the Onondaga Formation and the peritidal and sabkha facies of undifferentiated Lucas Formation in Michigan are correlative. Brett et al. (2011) proposed alternative explanations for such lateral lithological patterns, suggesting that
the top of the undifferentiated Lucas Formation should all be placed below the base of Onondaga Formation, and the Detroit River Group regarded as Lower Devonian deposits that are older than the Onondaga Formation (and thus equivalent to the Bois Blanc and Schoharie Formations in New York). They have also inferred that there was a major hiatus between the Lucas Formation and the Dundee Formation, when the Onondaga Formation was deposited in New York. In this correlation scheme, the Anderdon Member represents a local lithofacies of the Moorehouse Member and is considered much younger than the Lucas Formation, therefore it should be a member within the Onondaga Formation instead of the Lucas Formation. Consequently, they have avoided correlating the shallowest Lucas Formation deposits with the deepest Nedrow Member shales, but raised more questions than answers in contrasts with almost all the previous biostratigraphic studies (Landes 1951; Fagerstrom 1966, 1971, 1982; Linsley and Kesling 1982; Ludvigsen 1986). Other evidence that disputes this correlation comes from Core 19 Consumers 33409, drilled 5 km north of Port Alma, in which the Lucas Formation comprises undifferentiated Lucas Formation lithofacies (interbeds of anhydrite nodules and dolo-laminites) and the Anderdon Member lithofacies (lagoonal wackestone with stromatoporoid and coral fragments), is interpreted as a cyclic lagoonal to peritidal-sabkha depositional environment — consistent with the member status of Anderdon within the Lucas Formation (see Birchard et al. 2004 for details). Therefore, it is likely that the sabkha and peritidal facies (undifferentiated Lucas Formation), lagoonal facies (Anderdon Member of the Lucas Formation) and the biostromal facies of the lower Moorehouse are synchronous deposits on a paleo-ramp that inclined towards the east into the Appalachian Basin centre.
Figure 7.6 Cross section of A-A’. For well locations, please refer to Figure 6.1. The lateral contact of the Amherstburg Formation and the Onondaga Formation is speculative.
7.9 Summary

Lithofacies distributions indicate that the Onondaga and Dundee formations in Niagara area represent marine sediments deposited on a carbonate ramp, and that the Onondaga Formation accumulated in shallower water setting than its counterpart in New York. The ramp was inclined eastward towards the Appalachian Basin axis (Cassa and Kissling 1982), and the tectonic migration of the forebulge region may have formed a barrier that separated the Dundee lithofacies restricted in western New York and Ontario from the upper Onondaga Facies in central and eastern New York (see Ver Straeten and Brett 2000). In general, the ramp sedimentation of the Onondaga Formation began in eastern New York where it appears to conformably overlie the Early Devonian Schoharie Formation (Lindholm 1969; Brett et al. 2011), and the deepest water deposits in central New York (Cassa and Kissling 1982; Brett and Ver Straeten 1994). On the Niagara Peninsula, Onondaga Formation deposition followed a hiatus represented by a sand unit above the Bois Blanc Formation. Slightly deeper ramp lithofacies of the Edgecliff Member in central New York grades westward into western New York and Ontario, and the linear belt of bioherms and the subsurface biostromal facies indicate an outer to middle ramp depositional environment that shoaled towards the west and northwest. The ramp model is also supported by the occurrence of discontinuous biothermal distributions and the concentric arrangement of the lithofacies belt along the farfield side of the Appalachian Basin axis (see Cassa and Kissling 1982; Wolosz 1992). In Ontario, the occurrence of carbonate intercalations of fragmented corals and brachiopods in the cherty coral-crinoidal wackestone-packstone matrix of the Edgecliff Member indicates a shallow, warm open marine environment with periodic storm disturbance.

In central New York, the Nedrow Member consists of calcareous shales and argillaceous limestone interbeds, representing basinal facies, with starved carbonate production (Cassa and Kissling 1982). In Ontario, shales in the Onondaga Formation are absent and the argillaceous lime
mudstone grades into the Clarence Member lithofacies that is typically sparsely fossiliferous, representing the deepest water deposits of the Onondaga Formation in the Niagara Peninsula. The fauna in the Moorehouse Member is dominated by brachiopods in central New York and tabulate and rugose corals in western New York (Wolosz 1991). In Ontario, these biostromal facies commonly display graded beds of fossiliferous wackestone and calcisiltites that suggest winnowing mud from the skeletal grains. The winnowed coarsely crystalline interbeds in the Moorehouse Member were possibly deposited at water depth above normal wave base and represent the shallowest deposits within the Onondaga Formation. The sandy base of the Dundee Formation that overlies the Moorehouse Member suggests a return to carbonate deposition following a depositional hiatus during middle Moorehouse time. Discovery of the Tioga ash bed within the upper Dundee Formation has restrained the chronologic framework of these two units on either side of the forebulge region, making it possible to correlate into Michigan, Ohio and Pennsylvania where the ash bed is present.
CHAPTER 8. Sequence Stratigraphy of the Onondaga Formation in Niagara area, southwestern Ontario

To solve the correlation problems of the Onondaga Formation in Ontario westward into the Michigan Basin and eastward into the Appalachian Basin, sequence stratigraphic methods have been employed in this study. The study of cyclicity in carbonate successions is important to correlate the Early-Middle Devonian depositional units from the Appalachian Basin into the Michigan Basin across SW Ontario. It is also significant to test the global sea-level changes vs. endemic sea-level changes in stable cratonic interiors (Michigan Basin) and continental margins (Appalachian Basin). Modern research on Devonian eustacy was first attempted by House (1983) who outlined some major global sea-level changes. Johnson et al. (1985) took critical steps towards defining twelve third-order sequences of the Devonian world, which have been grouped into two second-order sequences within Sloss’s (1963) subdivided Kaskaskia Megasequence in North America (Fig 8.1). The recently published global onlap curves by Haq and Schutter (2008) is largely based on Johnson et al. (1985) study for the Devonian, though no biostratigraphic data was corroborated. Brett (1995, 1998) and Catuneanu (2002) have applied biostratigraphic and chemostratigraphic data into sequence stratigraphy and attempted to define sequences in North America with specific systems tracts. Brett et al. (2011) has established a revised sequence stratigraphic framework of the Devonian depositional systems in eastern North America, and this study is largely based on this work with minor revisions.

The cyclic Lower-Middle Devonian carbonate successions formed two third-order sequences based on the sequence stratigraphic subdivision scheme in North America (Sloss 1963; Johnson et al. 1992; Brett et al. 2011), which can be placed as Ic and Id defined by Johnson et al. (1985) (Fig
The Ic followed a global sea lowstand. In Ontario, its lower boundary is represented by the sporadic patches of siliciclastic sandstones (Oriskany and Springvale) on an Early Devonian karstic top of the Bois Blanc Formation (Johnson et al. 1992). The Ic sequence has a base defined by the first occurrence of conodont \textit{P. patulus/partitus} and a top by the first occurrence of \textit{P. australis}. It crosses the boundary of Emsian and Eifelian (Early Devonian-Middle Devonian) indicated by Johnson et al. (1985). Both the Detroit River Group and the Onondaga Formation form part of this third-order sequence (Johnson et al. 1992). Id is a middle Eifelian, third-order sequence with a top placed in the middle \textit{P. kockelianus} conodont biozone. The Dundee Formation is considered to comprise an overall transgressive system tract (Birchard 1993). In previous studies, limited work has been done to segregate each Devonian unit into systems tracts within each sequence. In addition, sequence stratigraphy is more difficult to apply in carbonate successions in craton interiors because the carbonate build-ups are controlled not only by eustatic sea-level change and basin subsidence, but also by climate change, paleogeography, tectono-sedimentary factors and carbonate factory productivity (e.g. Schlager 2005). This section is a first attempt to establish the sequence stratigraphic framework of the Onondaga Formation in Ontario and its correlation to the equivalent Devonian units in the Michigan Basin.
Figure 8.1 Devonian sea-level curve by Johnson et al. (1985). Study interval is in red box. Note the sequence Id started above the *P. costatus* costatus biozone.
8.1 Carbonate Sequence Stratigraphic Model

Wheeler (1958, 1963) was the first to systemically combine the methods of biostratigraphic, lithostratigraphic and sedimentologic methods to suggest the concept of time-units in the study of stratigraphy. The subject of sequence stratigraphy was established on the studies of the time-unit within each depositional system incorporated with seismic data (Vail 1987; Van Wagoner et al. 1988).

Siliciclastic stratigraphic sequences are defined as "a relatively conformable succession of genetically related strata bounded at their upper surface and base by unconformities and their correlative conformities" (Vail et al. 1974). The major control on the formation of siliciclastic system tracts is relative sea-level changes determined by eustatic sea-level changes and basin subsidence rates. The best-established global correlative frameworks are third-order sequences of variable duration (1-10 million years), though local tectonic activities may have complicated this comparison (Van Wagoner et al. 1990). High-resolution sequences (4th to 5th) sequences can only be correlated endemically, but is important for local sequence stratigraphic framework reconstruction.

The study of carbonate sequence stratigraphy differs from the siliciclastic sequences because the control on carbonate formation is governed by in situ biological carbonate factory in shallow to intermediate depth shelves and epicontinental seaways (Sarg 1988; Schlager 2005). Siliciclastic materials are transported from outside the basin, thus its accumulation is governed by relative sea-level rise to create physical accommodations. Rates of carbonate production, however, are greatest close to the water-air interface where photosynthetic processes are more active and nutrient level is high for calcite-secreting organisms. Moderate rates of relative sea-level rise allow the carbonate factory to produce carbonate grains at a much higher rate. Consequently, carbonate sequences are controlled by the biology and chemistry of the water realm known as “ecological accommodation”
There are four major variables controlling the geometry pattern of the carbonate sequence frameworks. They include 1) tectonic subsidence; 2) eustatic change; 3) production rate of carbonate factory and 4) climate (Sarg 1988).

Within a third-order sequence, two types of unconformities are defined as sequence boundaries: subaerial exposure surfaces (Type I) and subaqueous non-deposition surfaces (Type II) (Van Wagoner et al. 1990; Catuneanu 2002). In carbonate successions, when subaerially exposed, carbonates are prone to dissolution. Consequently, subaerial deposits (including tufa, calcrete, caliche deposits and paleokarstic systems, etc.) are good indicators of the sequence boundary Type I (Sarg 1988; Schlager 2005). A Type 2 sequence boundary is interpreted to have formed when the eustatic fall is less than the rate of basin subsidence at the carbonate platform margin, marked by a subaqueous non-deposition surface that is overlain by transgressive deposits (Sarg 1988). Above the Type II boundary, an aggrading shelf margin wedge is formed at the platform slope.

Specific carbonate sequence contains four depositional systems tracts are developed within a phase of relative sea-level change, including: lowstand systems tract (LST) or shelf margin wedge (SMW), transgressive systems tract (TST), highstand systems tract (HST) and falling-stage systems tract (FST) (Sarg 1988; Schlager 2005).

In siliciclastic sequences, a complete Type I sequence consists of LST-TST-HST-FST with a basal subaerial unconformity, and a complete Type II consists of SMW-TST-HST-FST overlying a submerged sequence boundary. In shallow marine and subtropical to tropical paleogeographic regions, the TST-HST are readily recognizable and the LST and FST are not always present. A sea-level drop may expose the shelf and an LST is established along the shoreline in the form of a lowstand fan or lowstand wedge deposited at the toe of the shelf-slope. In warm shallow marine
carbonate depositional environments, TST usually commence with shallow water facies, such as skeletal grainstone-packstone (shoal facies) or well sorted sandstones overlying an unconformity (Brett et al. 2011). During the HST, sea-level reaches its maximum and carbonate sediments are produced and form aggradating and progradating pattern seawards, giving an offlapping sediment package (Posamentier and Vail 1988). The maximum flooding surface between the TST and HST is usually marked by maximum starvation with deposition of condensed shale layers. In a type II sequence, the drop of sea-level rate is less so the shoreline migrates towards the shelf margin and a wedge developed there during lowstand (Van Wagoner et al. 1988).

In the carbonate platform/shelf slope, LST and SMW consist of a suite of prograding depositional systems when relative sea level falls (e.g. below or slightly above the shelf margin). TST is composed of a set of backstepping/retrogradational units that thicken shelfward when relative sea level rises above the previous shelf margin. TST onlaps landward and the younger units are commonly thinner because of limited accommodation (Catuneanu 2002; Coe 2003; Emery and Myers 2006). Unlike the siliciclastic TST, carbonate sequences usually contain thick sections of TST because the carbonate production often keeps pace with relative sea-level rise. Therefore, a thick retrogradational stacking geometry characterize the carbonate TST. Carbonate HST consists of the depositional systems formed during high sea level stands, and the facies belts prograde seaward that offlaps the maximum flooding surface (Schlager 2005). Carbonate HST is relatively thinner, because much of the accommodation space generated during transgressive time has already been filled simultaneously. In addition, rapid relative sea-level rise may cause a cessation of carbonate production, leading formation of condensed sections or hardgrounds with common pyritization. Carbonate FST is thought as significant sediment record formed when relative sea-level falls. Retreating sea has left sediment accumulation downward shift of the
shoreline and shelf surface (Hunt and Tucker 1992; Schlager 2005). These trace of FST forms a progradational geometry pattern seaward while subaerial conditions expose the platform margins.

The deposition of the Onondaga Formation in interpreted as on an easterly inclined ramp. In carbonate ramp depositional systems, the two important types are homoclinal ramp with little topographic and gradient change from shallow to deep, and the distally-steepened ramp with a break-of-slope at the deep end (Ahr 1973; Burchette and Wright 1992). Ramp system model by Burchette and Wright (1992) is divided into three parts, based on relative proximity to paleoshorelines: inner ramp, mid-ramp and outer ramp.

Inner ramp is characterized by sedimentation above the fair-weather wave base (FWWB) and may include carbonate facies of sabkha, lagoon, shoal, bank barrier and patch reefs (Flugel 2005). Type I sequence boundary is commonly well developed in inner ramp. On the mid-ramp, carbonate sedimentation usually takes place in the subtidal zone near the FWWB and above storm wave base (SWB), where skeletal wackestones/packstone/grainstones are commonly deposited with storm bed intercalations. Sequences developed on mid-ramps are dominated by TST and HST deposits, with LST deposited at the far end of the outer ramp (Schlager 2005). Complete sequence of LST-TST-HST-FST is commonly formed in outer ramp, and condensed sections may by marked by shale deposits or hardgrounds representing a shut-down of carbonate factory.

The general absence of topography on the homoclinal ramp precluded the generation of lowstand fans and wedges, unless disturbed by local tectonic uplift. Distally steepened ramps, have the potential for the formation of LST wedges during a lowstand, especially during a substantial sea-level drop down (Tucker et al. 1993).
8.2 Sequence Stratigraphic Distribution

As discussed previously, the Onondaga Formation was deposited on an easterly inclined ramp on the farfield side of the episodically active the Appalachian Foreland Basin centre region. Brett et al. (2011) has subdivided the Onondaga Formation into two third-order sequences in New York. The Edgecliff-Nedrow/Clarence-lower Moorehouse comprise the lower sequence that is comparable to Ic in Johnson et al. (1985), and the upper Moorehouse-Seneca (Dundee equivalence) form the TST of the Id sequence. In Ontario and western New York, the unnamed sandstone below the Onondaga Formation and the cherty fossiliferous limestone of lower Edgecliff Member was not deposited in central New York. We propose that the unnamed sandstone-lower Edgecliff represents the LST in the Ic sequence and the Edgecliff-Nedrow/Clarence-lower Moorehouse represents its overlying TST-HST.

8.2.1 Lowstand Systems Tract

In the western New York and Ontario, LST consists of one succession of siliciclastic sandstone overlain by restricted shallow marine limestone (Fig 8.2). In central New York, no carbonate deposition occurred due to the uplift of the forebulge region (cf. Ver Straeten and Brett 2000). The unnamed sandstone is thought to have formed at the slope of the forebulge region by reworked sands of Oriskany Formation, which represents a lowstand systems tract in western New York and Ontario. During the sea-level lowstands, the quartz sands infilled the paleokarstic surface of the Bois Blanc Formation, as seen in the form of thin neptunian dykes and fissure fillings (Cassa and Kissling 1982; Brett et al. 1994). This sandstone unit is missing in the central New York region, where a combined unconformity underlying the lower Springvale Member is formed (Brett et al. 2000). This siliciclastic sand unit grades westward into the lowermost cherty carbonate mudstone of the Amherstburg Formation as seen in the Core 11 (OGS 82-3) and Core 9 (Consumers’ Amoco...
Figure 8.2 Depositional scenario of the unnamed sandstone and the lower Edgecliff Member of the Onondaga Formation (LST). The forebulge region is interpreted to have been located east of Buffalo (Ver Straeten and Brett 2000). The unnamed sandstone is distributed west of the forebulge region and is absent west of Core 3 (Consumers’ Pan Am 13058). The exact lateral extent of this sandstone unit in the subsurface is unknown. This sandstone may have been from the sands of Oriskany Formation in response to the uplift to the east. Prior to the deposition of the Onondaga Formation, the short-lived uplift episode may have led to the erosion of the Lower Devonian, including the Oriskany Formation, in central New York. The lower Edgecliff Member is interpreted to have formed in a restricted shallow marine environment that was possibly connected to the Laurentia craton interior and disconnected from the central New York. During the deposition of the cherty, fossiliferous limestone of the lower Edgecliff, no deposition occurred in central New York. The contact of the Edgecliff and the Amherstburg Formation is not exposed in outcrop or recognized in any cored wells studies. This contact is thought as gradational in the zone between Core 10 (Consumers’ Pan Am 13057) and Core 11 (OGS 82-3). Please refer to Table 1.1 for well names and Figure 6.1 for well locations.
13076), in which the Bois Blanc-Amherstburg contact is marked by a thin layer (3-5 cm thick) of argillaceous quartz arenite. Together, the unnamed sandstone unit below the Onondaga Formation and the lowermost Amherstburg Formation comprise a siliciclastic-carbonate suite that features the lowstand wedges above the basal unconformity.

The overlying light brown, cherty, highly bioturbated coral-brachiopod wackestone-packstone of the lower Edgecliff Member in Ontario shows a prograding pattern westward into the Michigan Basin. It is capped by the top subaerial erosion with higher argillaceous materials in Niagara. Because of the lowstand of sea-level, the higher water energy has disturbed the skeletal fossil grains to form several storm beds in the lower Edgecliff Member. In Core 11 (OGS 82-3) and Core 9 (Consumers’ Amoco 13076), the uppermost of the lower Edgecliff is marked by 5-10 cm thick, light bluish grey, crinoid-brachiopod grainstone, indicating a period of winnowing and reworking. During the later transgressive stage, winnowed lag deposits of skeletal grainstone are locally developed that marks the initiation of TST (e.g. Core 10 Consumers’ Pan Am 13057).

8.2.2 Transgressive Systems Tract

The basal initial transgressive surface is marked by the occurrence of large crinoid columnals in Niagara area and New York, and a crinoid-rich packstone-wackestone facies above the light grey, pelletal grainstone in Ontario. This contact is not always easily recognized in cored wells, because the gamma ray log profiles may not show a positive excursion across this contact where no glauconites are present.

In Ontario, the crinoid packstones of the Edgecliff Member clearly onlap a subaerial surface, forming the early transgressive systems tract (ETST) (Fig 8.3). Where the lower Edgecliff Member is absent (e.g. in Ridgemount Quarry South and Norfolk Quarry), packstone with large crinoid fossils overlying the unconformity above the Bois Blanc Formation. It is inferred that the water
transgressed westward from the Appalachian Basin onto the Laurentian cratonic region (farfield side of Appalachian Foreland Basin). In the middle Amherstburg Formation in Ontario, a crinoid-rich packstone can be correlated with crinoid packstones of the Edgecliff Member (e.g. Core 11 OGS 82-3; Core 19 OGS 82-2). The crinoid-packstone lithofacies provided a substrate for bioherm initiation. The crinoid packstone facies may have thickened and thinned due to the irregular paleotopography of the sea floor – and possible responses to short-lived tectophases and karstification of early lithified skeletal sands. For example, in Core 2 (U.S Steel No.1), this unit is 3.5 m thick and in Core 3 (22 km southwest of Core 2) it is only 70 cm thick. Channel deposits may exist by the evidence of less fossiliferous crinoid mudstone-wackestone in Core 10 (Consumers’ Pan Am 13057).

The upper Edgecliff records deposition in an overall late transgressive systems tract (LTST). The pervasive depositional conditions were characterized by open marine waters that favoured the establishment of biostromal and biohermal complexes. Small patch reefs/bioherms created an irregular topography during this time, and biostromes were established between bioherm complexes (Fig 8.4). In the Amherstburg Formation in Core 11 (OGS 82-3) and Core 9 (Consumers’ Amoco 13076), an *Amphipora*-rich packstone intervals occur above the crinoid packstone, which represents a lagoonal facies at the back of the bioherm-biostromal build-ups in Niagara area. The bioherms-biostromes started to build up corresponding to the rising sea level. Several cycles of fossiliferous wackestone to mudstone have been recognized in cored wells, indicating a periodic transgression of the sea level. The cycles display distinctively different motifs. In Core 2 (U.S. Steel No.1), only one cycle can be recognized. Moving westward, the upper Edgecliff is thicker, and four shallowing-upward parasequences have been identified in Core 10 (Consumers’ Pan Am 13057). The parasequences are dominated by sparsely fossiliferous crinoid-coral wackestones
capped by thinly bedded, crinoidal packstone intervals. These are interpreted as developed within
the LTST stage. Each of the parasequences represents a somewhat asymmetrical, shallowing
upward succession capped by crinoid-rich shoal facies, abruptly overlain by deeper water
fossiliferous wackestones. Lenticular and tabulate stromatoporoids start to occur at the top of the
Edgecliff Member. Therefore, the top of the Amherstburg, as seen in Core 11 (OGS 82-3) and Core
9 (Consumers’ Amoco 13076), is likely to represent the same LTST stage with the upper Edgecliff
Member in Niagara area.

The top of the cherty mudstone of the Clarence member above the cessation of bioherm-
biostrome build-ups marks the maximum flooding surface, which is in turn overlain by the
fossiliferous limestone of the Moorehouse Member. In New York, the shales in Nedrow Member
may represent the deepest water deposits in the Onondaga Formation (Brett et al. 2011). Where
bioherms are present, this Clarence Member is very thin to absent, and the maximum flooding
surface is marked by the Edgecliff-Moorehouse contact.

The reef/biostrome facies at the top of the Amherstburg Formation in Core 11 (OGS 82-3)
and Core 9 (Consumers’ Amoco 13076), is overlain by another Amphipora-rich lagoonal facies
that is possibly correlative to the Clarence Member. However, the reef/biostrome growth resumed
above this lagoonal facies, indicating its growing continuity into the HST stage.

The deepening may be associated with the Chotec Bioevent of the early Eifelian (patulus to
costatus Zones; Klapper 2009). This transgressive systems tract, represented by abundant corals
and crinoids, could thus be correlated with the coral-rich members in Columbus Formation (Units
C and D) described by Staufer (1909) or the Bellepoint Member of the Columbus Formation in
Ohio (Judge 1998). In general, a global eustatic sea-level rise may have caused the deepening of
the epeiorogenic seas and marginal seas in Ontario with short-lived tecto-phases of the forebulge.
Figure 8.3 Depositional scenario of the middle Edgecliff Member and the lower Amherstburg Formation (ETST). As the sea level rises, the Laurentia craton interior was thought to connect to the Appalachian Foreland Basin. The occurrence of the crinoid-packstone to grainstone facies represents the initial transgressive surface. It has a regional extensive distribution that can be traced southward into Sarnia area in the subsurface. The lateral contact of the Edgecliff Member and the Amherstburg Formation is unknown.
Figure 8.4 Depositional scenario of the upper Edgecliff Member and upper Amherstburg Formation (LTST). As the sea level continues to rise, bioherms/biostromes are well developed in the upper Edgecliff Member. Based on the fauna studies, the bioherms in Edgecliff Member may have developed in a subtropical environment or a water depth near the base of the photic zone. No stromatoporoids or massive tabulate corals are found as bioherm frame-builders. Stromatoporoids started to occur in Core 10 (Consumers’ Pan Am 13057). In Core 11 (OGS 82-1) and Core 9 (Consumers’ Amoco 13076), the top of the Amherstburg Formation is featured as abundant stromatoporoid-tabulate corals reefal facies. It is likely to have formed in a shallow warm water, high energy environment. Therefore, the interpreted ramp sloped to the east and the Amherstburg represents a shallower water facies than its counterpart of the upper Edgecliff Member in Niagara area and western New York. Bioherm/reef is not to scale and has been enlarged for illustrative purposes.
8.2.3 Highstand Systems Tract

The Nedrow/Clarence-Moorehouse Member contact is widely spread in New York and Ontario (Brett et al. 2011), representing a return to shallow, open marine carbonate depositional environment (Fig 8.5). Above the maximum flooding surface, faunal diversity is lower than that in the Edgecliff Member, which represents a recovery of faunal communities. It consists of echinoderm grains and low diversity rugose corals, suggesting benthic conditions that was possibly unfavorable to many other benthic organisms (Fig 8.6). Chert nodules are still common with poor recovery of brachiopods. Above the base of the cherty interval, a general shallowing upward succession is observed in cored wells. For example, three shallowing-upward cycles have been recognized in Core 5 (Consumers’ Amoco 13102). Each cycle consists of the lower non-cherty coral-crinoid wackestone and an upper light grey, crinoid-coral grainstone. These cycles represent intervals of shallowing up-section, from shallow marine environment to crinoid shoal environment. Locally, the lower fossiliferous wackestone is featured as biostromal facies with abundant *Eridophyllum* and *Heliophyllum*. In Core 10 (Consumers’ Pan Am 13057), tabulate and lenticular stromatoporoids are abundant, which resembles the biostromal facies at the top of the Amherstburg Formation in Ontario. It is thus inferred that the Amherstburg Formation reefs/biostromes have thrived into the lower Moorehouse time. Fagerstrom (1961, 1983) suggested that there are two cycles in the reef growth of the Amherstburg Formation. The lower cycle consists of stromatoporoid-coral bafflestone encrusted by *Amphipora*-rich or pelloital lagoonal intervals. The upper cycle is featured as less diverse stromatoporoid-coral bafflestone capped by a light grey, leached reef top. Tracing laterally, Fagerstrom (1983) suggested the lower cycle represents an aggradation and the upper cycle progrades basinward. In our study, evidence to support the growth cycles comes from Core 11 (OGS 82-3) and Core 9 (Consumers’ Amoco 13076), in which the reefal/biostromal facies at the top of Amherstburg Formation consists of two similar cycles. Each
cycle is capped by tan to light grey, leached *Amphipora*-rich grainstone that marks the cessation of reef growth. It is thus speculated that the lower aggradation cycle represents the “keep-up” pattern in LTST and the upper cycle forms a “give-up” pattern at the early stage of HST. These two cycles are separated by the maximum flooding surface traced into Clarence/Nedrow Member.

The late highstand to fall-stage systems tract is represented by the crinoid-brachiopod grainstone (shoal facies) at the top of the lower Moorehouse Member (Fig 8.7). The winnowed skeletal grains are interpreted as having been transported down ramp from the west.

The Anderdon Member of the Lucas Formation in Ontario is also regarded as representing a highstand systems tract. Anderdon Member, however, represents a back-reef facies or lagoonal facies with very common stromatoporoids, corals and *Amphipora*. It may represent the cratonic interior counterpart of the Moorehouse Member in Ontario. The relationships between the Anderdon Member and the Moorehouse Member is seen in Core 10 (Consumers’ Pan Am 13057). In this core, a 1.5 m thick, buff to tan, very finely crystalline, sparsely fossiliferous Anderdon Member overlies the Moorehouse Member by a gradational lithological change zone. West of this core, in Core 11 (OGS 82-3) and Core 9 (Consumers’ Amoco 13076), the Anderdon Member comprises the entire Lucas Formation, and the Moorehouse Member is not recognizable. It is interpreted that during the HST, the lagoonal facies of the Anderdon Member offlapped the top of the Amherstburg Formation and the Moorehouse Member in Ontario, forming an eastward prograding wedge.
Figure 8.5 Depositional scenario of the Clarence/Nedrow Formation and the upper Amherstburg Formation (EHST). The sea-level reaches its maximum, and the shales of the Nedrow Member started to deposit in central and eastern New York. Rise of the sea level has caused the cessation of biohermal/reefal growth. Deep water facies of cherty mudstone was deposited above the biostromal facies of the upper Edgecliff Member and an *Amphipora*-rich packstone-grainstone interval caps the reefal facies at the top of the Amherstburg Formation.
Figure 8.6 Depositional scenario of the lower Moorehouse Member and the Anderdon Member of the Lucas Formation (EHST). As the sea-level remains high, the biostromal facies recur in the lower Moorehouse Member. The Anderdon Member in core 11 (OGS 82-1) and Core 9 (Consumers’ Amoco 13076) comprises of the whole Lucas Formation. It is featured as back-reef or lagoonal facies of coral-brachiopod-*Amphipora* wackestone facies with locally developed biostromal facies.
Figure 8.7 Depositional scenario of the lower Moorehouse Member and the Anderdon Member of the Lucas Formation (HST-FST). In the HST-FST stage, the Anderdon Member onlaps the Moorehouse Member as seen in Core 10 (Consumers’ Pan Am 13057). This contact indicates that the Anderdon Member is at least partly equivalent to the lower Moorehouse Member. In Niagara and New York, the top of the Moorehouse Member is capped by crinoid-brachiopod grainstone that represents a shoal facies. An unconformity is regionally extensive above the Moorehouse Member and the Anderdon Member, which represents the upper sequence boundary.
8.2.4 Dundee–upper Moorehouse–Seneca Cycle

The succeeding cycle (Id) started from the base of the Dundee Formation in Ontario and middle Moorehouse Member in New York above a regionally extensive unconformity. Johnson et al. (1985) curve shows that the Id cycle started above the *P. costatus costatus* zone, however, the Dundee Formation and the upper Moorehouse Member are all placed below the last occurrence of *P. costatus costatus*. This conflicts with Johnson et al. (1985) curve, and Brett et al. (2011) suggested a revision of the biostratigraphic zones in Johnson et al. (1985) curve.

This sequence is bounded by a lower Type I sequence boundary. In Ontario, the upper contact of the Anderdon Member of the Lucas Formation is marked by a regional unconformity below the Dundee Formation (Uyeno et al. 1982). Reworked quartz grains in the basal Dundee Formation indicate a long period of non-deposition prior to the deposition of Dundee Formation. In central-eastern New York, this unconformity in the middle Moorehouse has been reported by Brett et al. (2011). It is thought to be continuous across the Appalachian Basin into the Michigan Basin, representing the basal sequence boundary of Id cycle. The Dundee Formation depositional history has been well summarized in Birchard (1993). A carbonate ramp model with low gradient slope from shallow water shoreline southwestward into lagoon and open marine depositional environment is proposed (cf. Read 1995).

In the Niagara area, it seems that the water transgressed from Appalachian Basin eastward into the craton interior and formed a prograding TST evidenced by the open marine lithofacies of the Moorehouse/Seneca Member into the lagoonal deposits of Dundee Formation. Within the Dundee Formation, two fourth-order cycles have been recognized. The first cycle consists of a basal skeletal carbonate sand base of the Dundee Formation deepening upwards into the very cherty carbonate mudstone facies (LF12). It is speculated that the waters between the craton interior and the foreland basin were not well interconnected and the paleotopographic highs on the
farfield side of the Appalachian Basin played a role in separating the two depositional realms in early Dundee deposition. The early transgressive stage deposits comprise the very cherty, fossiliferous limestone with very common argillaceous seams (LF10-11). The overlying shallow marine facies (LF11) indicates a relative sea-level rise. Later transgression formed a dysaerobic, lagoonal mudstone facies (LF12), and an episode of relative sea stillstand has led to various intervals of firmground formation (Birchard 1993). The presence of *Tasmanites* and a high burial rate of organic matter has contributed to the formation of a source rock for oil and gas (Liberty and Bolton 1971). This lagoonal facies is overlain by either a regionally extensive firmground (Birchard 1990) in Michigan Basin or brachiopod-crinoid grainstone in Niagara area, indicating an episode of winnowing and reworking. It is in turn overlain by the brachiopod wackestone-packstone by a sharp contact, and the lithologies become muddier up-section. This shift represents another deepening stage evidenced by the absence of chert nodules, the muddier lithology and the fewer grainstone pulses within LF13. As the sea became deeper, the two basins were likely to have been well connected in Ontario, evidenced by the occurrence of the same lithofacies of argillaceous brachiopod mudstone-wackestone in upper Dundee Formation and in Seneca Member (Brett et al. 1994). The uppermost Dundee Formation is overlain by the shales of Marcellus Formation or Hamilton Group that represents a maximum flooding surface within the Id cycle. In general, the Dundee Formation may have been formed during a transgressive condition with periodic shoalings and seawater stillstands, and the two basins were well connected at the end of the Dundee deposits.

**8.3 Summary and Future Work**

In general, the Onondaga Formation represents a carbonate ramp succession in Ontario and western New York that grades into the Amherstburg, Lucas and Dundee Formations westward.
The argillaceous limestones (e.g. the Nedrow and lower Moorehouse) are confined to the basin axis in central New York. In Ontario, the Onondaga Formation lithofacies grade into biostromal and cherty limestones that represent a shallower water, higher energy ramp setting, and its cherty wackestone, cherty mudstone to biostromal packstone-grainstone lithofacies mark a succession from slightly deeper water low-energy environments (Clarence Member) to higher energy shoals into the lower-middle Moorehouse Member of the Onondaga Formation. The occurrence of the Tioga ash beds within the Dundee Formation may provide a regional correlative marker bed for its correlation into the upper Onondaga Formation in New York and the contact of Columbus and Delaware formations in Ohio.

The unnamed sandstone-Edgecliff-Nedrow/Clarence-lower Moorehouse defines a third-order sequence in Ontario. The unnamed sandstone below the Onondaga Formation and the cherty fossiliferous lithofacies of the lower Edgecliff in Ontario represent the lowstand systems tract, which is absent in New York. It may be correlated to the sandy top of the Schoharie Formation in eastern New York. Distribution of the sandstone was controlled by the migration of the forebulge region that resulted in the reworking and redistribution of the Early Devonian sandstone units (Oriskany Formation). Widely distributed crinoid grainstone-packstone formed during the initial transgressive stage of Ic. It is traceable and correlative well into the middle Amherstburg Formation in Ontario and Michigan. Bioherms and biostromes were formed in the late transgressive systems tract, creating an irregular paleotopography. The top of the Nedrow/Clarence Member represents the maximum flooding surface, marked by the occurrence of shales or very cherty mudstone. The two cycles of reef/biostrome growth at the top of the Amherstburg Formation may cross the maximum flooding surface into the high system tracts, indicating a “keep-up” to “give-up” growth pattern. It is thus inferred that the Anderdon Member of the Lucas
Formation and the Moorehouse Member are correlative and form a prograding pattern from Niagara area into the Appalachian Basin center eastward.

However, the sequence stratigraphic model is largely based on core logging, and there are still gaps between cores where no data was collected. For example, the drastic lithofacies change occurs between Core 10 (Consumers’ Pan Am 13057) and Core 11 (OGS 82-3). No data has been gathered between these two cores. More drilling is needed to incorporate geophysical and cutting data, and to establish the exact contact relationships of the Onondaga Formation and the Detroit River Group.
CHAPTER 9. Conclusions

This study forms a part of the regional-scale bedrock groundwater mapping initiative within the Groundwater Program of the Ontario Geological Survey. Results are presented as a part of this multi-year bedrock study to map the upper Silurian–Middle Devonian sequence stratigraphy based on logging and sampling of 38 continuously cored holes across SW Ontario. The stratigraphic units investigated in this thesis include the Silurian Bass Islands/Bertie formations and the Devonian Onondaga Formation. The S-D unconformity in SW Ontario has been well documented. Lithofacies characterization has been carried out within a sequence stratigraphic framework across the Lake Erie Region of the Appalachian Foreland Basin and across the cratonic forebulge region (eastern Michigan Basin), with emphasis on the depositional breaks and associated bedrock groundwater flow zones that reside in the upper Silurian through Middle Devonian strata of SW Ontario. Recognition and correlation of these units and unconformities is important in the interpretation of tectophase durations, sea level fluctuations, paleogeographic configurations and paleoclimatology.

General results are presented as follows.

1. The Bass Islands and Bertie formations are the youngest Silurian deposits in SW Ontario. The Bass Islands/Bertie formations overlie the evaporite-carbonate succession of Salina Group and underlie the Devonian Oriskany Formation siliciclastic sandstones and/or the Bois Blanc Formation cherty carbonates. Twelve lithofacies have been recognized within the Bass Islands Formation in cored wells. The peritidal-sabkha dominated evaporite-carbonate successions indicate a restricted, semi-arid and hypersaline environment when the Bass Islands Formation was
deposited. Based on the lateral distribution of the lithofacies, it is inferred that the Michigan Basin depositional realm in Ontario indicates a widespread, restricted and evaporitic craton interior and the Appalachian Foreland Basin depositional realm in Niagara area represents an inner ramp environment. The open marine lithofacies in the Niagara area and the restricted lithofacies to the west suggest a possible easterly inclined ramp depositional environment across SW Ontario. Stratigraphic correlation between each unit is difficult because they were deposited in very shallow waters. A minor relative sea level change may cause drastic lithofacies change both laterally and vertically.

In this study, three depositional cycles were recognized within the depositional succession of Salina G Unit to the Bass Islands Formation. The lower Cycle 1 consists of a 3–5 m thick, regionally traceable shallowing-up succession. It characterizes a change from regional restricted, possible stagnant and anoxic environment to later peritidal-sabkha environment. Cycle 2 is featured as several cycles of subtidal-supratidal deposits with periodic subaerial exposures. Cycle 3 represents shallow subtidal, well circulated water deposits formed prior to the S-D unconformity.

2. Stratigraphic correlation between the Bass Islands Formation and Bertie Formation has been revised. The five members in Bertie Formation, in descending order, are: Akron, Williamsville, Scajaquada, Falkirk and Oatka. All these five members of Bertie Formation are correlated to the upper Salina F through to the lower Bass Islands Formation. Based on outcrop section observations, the Bass Islands Formation overlies the Akron Member in Dunville, Cayuga and Hagarsville quarries. It is inferred then the Bertie Formation is older than the Bass Islands Formation. The Akron Member represents a restricted, hypersaline environment that is correlative to the evaporitic dolomites of “False G Unit” in the lower Bass Islands Formation. The Williamsville and
Scajaquada members were assigned to Cycle A at the base of Bass Islands Formation. The Falkirk Member is lithologically the same as the lower Salina G Unit, and the green shale of the Oatka Member shows the same mineralogical feature as the Salina F. All members of Bertie Formation contain no evaporite.

3. The Bass Islands Formation displays a variety of paleokarst features during and after its deposition below the S-D unconformity. Two possible traceable paleokarstic profiles are documented. The lower paleo-cave systems are formed at the base of Cycle 2 above the top of the evaporitic “False G Unit” or Akron Member. It is speculated that the caves formed syn-depositionally or during an early diagenetic stage. Dissolution-collapse breccias are readily recognized. Cave floor, cave infill and cave roof could thus be identified due to their distinctive brecciated features. The upper regional extensive paleokarst system is right below the S-D contact. Various paleokarstic features are well developed to form regional extensive bedrock aquifers.

4. The paleokarst profiles in the Bass Islands Formation have potentials to form bedrock aquifers in SW Ontario. Four paleokarst events are recognizable: 1) Evaporite dissolution. Dissolution of the evaporites in Salina G and the lower Bass Islands Formation may comprise multiple stages of dissolution-collapse breccias. The dissolution time is hard to determine. It may have taken place right after the evaporite deposition or as late as during Acadian Orogeny in Middle Devonian time. 2) Syn-depositional dissolution. Brecciation occurred during the periodic subaerial exposure. Complex dissolution fabrics including desiccation cracks, tepees, caliches, dewatering structures and collapse breccias were formed in the vadose zone above the water table. These profiles may have penetrated several pre-existing carbonate profiles and made the correlation between each unit
difficult. 3). Pressure-dissolution breccias. When the Bass Islands Formation was deeply buried, peaked stylolites containing large amount of clay impurities may have formed. This is evidenced by the clast-supported breccias surrounded by peaked argillaceous stylolites in the lower Bass Islands Formation (including the Falkirk Member). 4). Surface paleokarst below the S-D contact. The regional uplift of SW Ontario during the late Silurian–Early Devonian has exposed much of the carbonate deposits. A complex paleokarst system was formed, featuring solution-widened joints, solution-vugs, leached patches, sinkholes, collapse breccias, and in places filled with variable siliciclastic materials. This event may have accentuated the previous karst profiles. The migration of water table and brine level is the key to distinguish the formational environment of different karst levels.

5. Three unconformities have been recognized in SW Ontario across the S-D contact, including, the Bass Islands-Oriskany, Oriskany-Springvale (Bois Blanc) and Bois Blanc-Onondaga unconformities. The Bass Islands-Oriskany hiatus may have coincided with the erosion of a substantial part of the upper Silurian strata and is marked by an initial stage of the deposition of Devonian sands. These sands are not continuously distributed on regional scale, being preferentially preserved in paleokarst sinkholes or solution-enhanced joints. The Springvale glauconitic and phosphatic quartz-rich sandstone mixed with carbonate minerals create a calcareous sandstone zone that marks a second unconformity. The unnamed sandstone below the Onondaga Formation, which is regarded as equivalent to the Sylvania Member in southern Ontario, marks the third unconformity. All these clean and well-sorted sands are interpreted as eolian in origin. The provenance is not determined, though it seems to have a mixed origin from the northeast and from the west.
6. The sandstone unit below the Onondaga and above the Bois Blanc Formation has been misidentified as the Springvale Member in previous studies. It should be separated from the Springvale Member (lower Bois Blanc Formation) in New York and Ontario. The occurrence of this sandstone unit is regionally recognizable between the underlying Bois Blanc Formation and the overlying Onondaga Formation. In New York, this sandstone unit may have amalgamated with the Springvale Member sandstone where the carbonate rocks of the Bois Blanc Formation are absent, making it not readily correlative in cored wells in Ontario.

7. The Onondaga Formation is the upper Lower Devonian–lower Middle Devonian unit in the Appalachian Foreland Basin. It is exposed and subcrops in the Niagara area and transitions laterally into the Detroit River Group (Amherstburg and Lucas formations). The nomenclature of “Onondaga Formation” was not adopted in previous well data in Ontario. The fossiliferous and dark grey cherty limestones of the Onondaga Formation represent an open marine depositional environment, which differs from the more restricted lagoonal or inland sea deposits of the Amherstburg Formation and the sabkha-dominated deposits of the Lucas Formation. Therefore, it is necessary to apply the name Onondaga Formation in the Niagara area. Only the lower three members of the Onondaga Formation are present in Ontario, which are, in ascending order: the Edgecliff, Clarence/Nedrow, and Moorehouse members.

8. The previously identified cherty limestone of the upper Bois Blanc Formation above the unnamed sandstone unit should be assigned to the lower Edgecliff Member based upon its Onondaga-aged megafauna. The lower Edgecliff Member resembles the lower Amherstburg Formation in its brownish grey, cherty and slightly argillaceous limestone lithology and similar
megafaunal assemblages, which represent an epicratonic or inner ramp depositional environment. Another possibility remains in that the lower Edgecliff Member may be, in places, an erosional residuum or erosional remnant of the Amherstburg Formation. These two carbonate units are impoverished in conodonts, making correlations more difficult. The lithofacies transition zone of the Onondaga and Amherstburg formations is tentatively located between Port Burwell and Port Bruce.

9. Based on the study of lithofacies distribution and interpretation, the Onondaga and Dundee formations in the Niagara area represent an easterly inclined carbonate ramp. In this study, six lithofacies were recognized in the Onondaga Formation and five in the Dundee Formation. The various Onondaga lithofacies suggest shallower water deposits than its counterpart in New York. The tectonic migration of the forebulge region may have formed a barrier that separated different lithofacies in Ontario and in central New York.

In the Niagara Peninsula, deposition of the Onondaga Formation followed a hiatus represented by a sand unit above the Bois Blanc Formation. Slightly deeper ramp lithofacies of the Edgecliff Member in central New York grades westward into western New York and Ontario, and the linear belt of bioherms and the subsurface biostromal facies indicate an outer to middle ramp depositional environment that shoaled towards the west and northwest. The shaly facies of the Onondaga Formation is absent and the argillaceous lime mudstone in central New York grades westward into the Clarence Member in Ontario that represents the deepest water deposits of the Onondaga Formation in the Niagara Peninsula. The Moorehouse Member represents a shallowing cycle capped by an unconformity with the overlying Dundee Formation. Discovery of the Tioga ash bed in the Dundee Formation has provided evidence for its correlation into New York. The ash
bed occurs 20–24 m above the base of the Dundee Formation. This ash bed marks the contact of the Moorehouse and Seneca members (Onondaga Formation) in New York and the contact of the Columbus and Delaware formations in Ohio.

10. A sequence stratigraphic framework has been established on the cored well data of the Onondaga Formation and Dundee Formation in Niagara area. The unnamed sandstone-Edgecliff-Nedrow/Clarence-lower Moorehouse constitutes a third-order sequence in Ontario, which can be placed as the Ic sequence in the Johnson et al. (1985) subdivision scheme. West of the forebulge region, the unnamed sandstone below the Onondaga Formation and the cherty fossiliferous lithofacies of the lower Edgecliff in Ontario represent the lowstand systems tracts, which is absent in New York. It may possibly be correlated to the sandy top of the Schoharie Formation in eastern New York. Widely distributed crinoid grainstone-packstone was deposited during the initial transgressive stage in Edgecliff Member. It is traceable and correlative well into the middle Amherstburg Formation in Ontario and Michigan. Bioherms and biostromes in the Edgecliff Member were formed in the late transgressive systems tract, creating an irregular paleotopography. The top of the Nedrow/Clarence Member represents the maximum flooding surface, marked by the occurrence of shales or very cherty mudstone. The biostrome facies in the Moorehouse Member consists of the HST, evidenced by a prograding eastward pattern.

11. The Amherstburg Formation and the Anderdon Member (Lucas Formation) in Core 11 (OGS 82-3) and Core 9 (Consumers’ Amoco 13076) show similar cyclicity to that in the Onondaga Formation in Niagara area. Conodont data also supports the stratigraphic correlation of the Amherstburg Formation and Anderdon Member with the Onondaga Formation. The Amherstburg
Formation and the Anderdon Member represent a lagoonal depositional environment compared to the open marine counterpart of the Onondaga Formation. This suggests that a paleotopographic highland may have separated these two depositional realms near Port Burwell.
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