STRATIGRAPHY OF THE MIDDLE TO UPPER ORDOVICIAN FORELAND-BASIN SUCCESSION, OTTAWA EMBAYMENT: EMPHASIS ON THE TURINIAN-CHATFIELDIAN SUCCESSION

BY

NKECHI E. ORUCHE
M.Sc. (University of Aberdeen)

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Carleton University
Ottawa-Carleton Geoscience Centre
Ottawa, Ontario

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Nkechi E. Oruche
ABSTRACT

New contributions are presented concerning the character of the depositional systems and stratigraphic architecture of the Middle to Upper Ordovician foreland-basin succession, and in particular the Turinian-Chatfieldian succession, that underlies the Ottawa Embayment, central-east Canada. The Ordovician foreland succession is the local expression of the cratonic Tippecanoe I Megasequence that extends across southern Laurentia, and developed within a retro-arc basin arising from distal plate-boundary convergence. This study investigates craton-interior intrabasinal response of sedimentation to tectonic, climatic, depositional and eustatic controls, and compares the basin’s stratigraphic fabric with equivalent strata along strike, to the south, in the much larger Appalachian Basin.

The embayment’s foreland stratigraphy is divided into platform and basin successions: in ascending order, seven (I-VII) transgressive-regressive (T-R) depositional sequences and one (VIII) transgressive sequence comprise a Middle to Upper Ordovician (Dariwillian to Edenian) platform succession whereas an incomplete regressive sequence (IX) forms the basin succession (Edenian through Richmondian). Combined, they define a tectonic cycle of net platform deepening over ~15 my and subsequent basin-fill spanning ~3 my. This asymmetry characterizes the local expression of diachronous foundering of the regional southern Laurentian margin related to orogen migration and erosion, then basin fill.

My study reaffirms placement of the Sauk-Tippecanoe I (S-T) Megasequence boundary within the upper Beekmantown Group: it is timed with significant synsedimentary structural events that appear to predate regional regression, then flooding, previously
used to define the S-T boundary at the top of the Beekmantown succession and its lateral equivalents in southern Laurentia. The majority of T-R depositional sequences within the embayment contain abrupt (often erosional) boundaries also best defined in terms of synsedimentary structural control, with a longer-term (Turinian-early Edenian) platform deepening timed with distal structural loading related to retro-arc shortening.

Examination of the upper Turinian-Chatfieldian (Upper Ordovician) succession in the embayment reveals three stages of carbonate-platform development of which the intermediate stage illustrates more intrabasinal (and structural) control on sedimentation compared to bounding stages that document regional continuity with extrabasinal stratigraphy:

a) *Stage 1* is represented by the Turinian Lowville Formation: its shallow-water carbonate facies is correlated inter-regionally suggesting a widespread depositional response to either eustatic or regional tectonic controls along southern Laurentia.

b) *Stage 2* incorporates the Watertown Formation, the newly defined L’Orignal Formation, and a newly defined coeval relationship of the Rockland Formation (revised) and hitherto unrecognized strata forming a revised lower Hull Formation. This stage illustrates marked intrabasinal differentiation of sedimentary facies and formation thicknesses as a result of lateral variation in differential subsidence. Depositional sequence attributes include prominent ravinement and erosional surfaces, and a post-Watertown record of seaward progradation of siliciclastics not obviously traced beyond the embayment’s limits. The Turinian-Chatfieldian boundary is defined at the top of the L’Orignal Formation by discovery of the Millbrig Bentonite in a quarry near L’Orignal (ON), in the eastern embayment. U-Pb isotope analysis of zircons has yielded the most
precise age date (453.36 ± 0.38 Ma) yet published for this ash-bed deposit that extends across the entire Appalachian Basin to the south. Immediately following this event, the once embayment-wide L'Orignal paleoplatform underwent segmentation into deeper (lower Hull Formation) and shallower (Rockland Formation) water settings.

c) Stage 3 marks re-establishment of inter-regional distribution of a common depositional facies characterized by progradation and expansion of high-energy crinoidal shoal systems of the upper Hull Formation traced into the Deschambault Formation in the Quebec Basin to the east, and Kings Falls Formation of the northern Appalachian Basin in New York. This change marks a regional shallowing, possibly of eustatic origin.

The upper Turinian-Chatfieldian interval was examined from the perspective of δ¹³C chemostratigraphy to provide higher resolution of intra- and interbasinal correlation. Four positive excursions (E1 to E4) are mapped through the Ottawa Embayment into two outliers to the northwest along the Ottawa-Bonnechere graben. Turinian excursions (E1 and E2) are lithostratigraphically constrained by cross-embayment erosional surfaces and coincide with periods of regional transgression. They can be correlated into sections underlying the Upper Mississippi Valley (central USA) illustrating their regional extent. Chatfieldian excursions (E3 and E4) occur in the coeval facies of Rockland and lower Hull formations, and upper Hull strata, respectively. Excursions E3 is the local expression of the regional (if not global) Guttenberg δ¹³C excursion (GICE). Of particular importance, improved resolution of the Guttenberg δ¹³C excursion in the embayment demonstrates examples of amalgamated (E3 and E4) excursions across disconformities. In combination with lateral variation in formation thickness and facies,
formation-specific δ¹³C excursions suggest local modulation of regional (or global) δ¹³C excursions in response to changes in circulation, productivity, and accumulation rates.

In summary, when compared to laterally equivalent sedimentary basins in southern Laurentia, the Middle to Upper Ordovician foreland succession of the Ottawa Embayment illustrates elevated syndepositional structural control on sedimentation. This is likely a far-field response to contemporary changes in subsidence rates and tectonic events along the distal convergent Laurentian plate, that were focused into the craton interior along an axis of inherited weakened continental crust defined by a late Precambrian intracratonic fault system that underlies the embayment. Tectonism played an important role in influencing higher order base-level changes along this part of the Laurentian platform that were superimposed on an otherwise net eustatic rise through the Middle and Late Ordovician.
ACKNOWLEDGEMENTS

First, I thank my supervisor, Prof. George Dix, who gave generous support to my research, both intellectually and financially through a Natural Sciences and Engineering Council Discovery Grant (NSERC). Your guidance, constructive discussions and position outside the box made this research to be of great interest. Financial support was also provided by Carleton University scholarship. I thank quarry owners (A.L. Blair Construction, Thomas Cavanagh Construction, Colcem Canada Ltd., H&H Construction) for granting access to their properties, and the Geological Survey of Canada (Ottawa) for access to cores. My appreciation is extended to departmental analytical support from Peter Jones (electron microprobe), Jianqun Wang (scanning electron microscopy), and Tim Mount (thin-section preparation). Thin sections were also prepared by Vancouver Petrographics Ltd (BC). Staff at the Jack Satterly Geochronology Laboratory (University of Toronto) are thanked for their support with stages of geochronological (U-Pb zircon) work. Carbon and oxygen isotope analyses were conducted at the Ján Veizer Isotope Laboratory (University of Ottawa) and Queen’s Facility for Isotope Research (Queen’s University, Ontario). As this dissertation contains three chapters submitted for publication in peer-review journals, I thank Profs. T. Algeo and F. Macdonald, and Dr. D. Lowe, as well as two anonymous reviewers, for critiques and suggestions that improved the scientific presentation.

I’d like to thank my office mate, He Kang, and my graduate student colleague and friend, Jade Atkins, for the advice, support and discussions.
To my parents, Prof. and Mrs. B. C. Egboka, who have been there from the beginning, your achievement in geology inspired me to follow your footsteps; and my in-laws, the late Mr. Ndukwu and Mrs. Mary Oruche, I say thank you for your love and for believing in me.

A special thanks goes to my husband, Kingsley Oruche, who supported me through many challenges and made even more sacrifices during the course of my study to allow me to pursue my dreams. Without his love, support, encouragement and inspiration, this project may never have been completed. Finally to our adorable angel, Lenore Oruche, thank you for your patience and how you are always happy to come with me to school.
STATEMENT OF ORIGINAL CONTRIBUTIONS

Research was conducted under the supervision of Prof. George R. Dix. The thesis is presented in a *three-paper format*; that is, the middle chapters (2-4) form peer-review article submissions, and are bound by chapters (1, 5) that introduce and summarize the research, respectively. Chapters 2 and 3 were published in *Canadian Journal of Earth Sciences* and *Palaeogeography, Palaeoclimatology, Palaeoecology*, respectively, and Chapter 4 was submitted (since submission of the thesis) to *Basin Research*.

I am the lead author on all three peer-review submissions and bounding chapters. This work represents a combination of field and laboratory analyses. I mapped rock exposures along roads, rivers and in quarries, and logged drill core; from thin sections, I conducted petrographic analysis; I was a lab technician involved in concentration and picking of zircons under the direction of Dr. Sandra Kamo (University of Toronto) for determination of a U-Pb age on bentonite (Chapter 2) at the University of Toronto; and, in another technical job, I was involved in the determination of carbon and oxygen isotopes (Chapter 3) at University of Ottawa. I drafted thesis art work, compiled data, and developed preliminary interpretations.

My supervisor provided scientific and editorial mentoring such that discussions about the science, presentation, and back-and-forth revisions of writing style resolved into manuscripts for submission. He compiled Figs. 4-5 and 4-6, and Appendix I, with their more paleontological focus. Remaining co-authors, Mr. Sean Gazdewich (Chapter 3), currently an M.Sc. student at University of Victoria, contributed data from his B.Sc. thesis (Carleton University); and, Dr. Sandra Kamo (University of Toronto) provided
technical language for description of U-Pb age dating reported in Chapter 2. All manuscripts were evaluated by my co-authors prior to submission. Journal reviewers provided additional recommendations in terms of re-organization, writing, and figure composition, all of which required rethinking and reformulation of some ideas and their presentation.

During the course of the research, preliminary results were presented at conferences for which I was the lead author (Oruche and Dix, 2016; Oruche et al., 2017). I was also a co-author on another conference presentation that used some of my data (Kang et al., 2017).

Significant scientific contributions arising from my own field and lab-based analyses:

In *Chapter 2*: (1) recognizing very prominent vertical and lateral facies divisions that illustrate greater variety in paleotopography and differential subsidence than recognized before; (2) extending the limit of the regional Millbrig K-bentonite into southern Canada, and reporting the most precise age date (through collaborative work with Dr. Sandra Kamo, University of Toronto) for this ash deposit that blanketed southern Laurentia;

In *Chapter 3*: (1) inter-regional correlation of Turinian δ13C excursions that had not been demonstrated previously; (2) improved resolution of the GICE that reveals stratigraphic amalgamation of excursions across disconformities; (3) the role of local oceanographic and depositional controls that modulate regional to global isotopic signals;

And, in *Chapter 4*, a significant new contribution to regional Ordovician stratigraphy in North America by demonstrating inter-regional discordance of sequence boundaries highlighting the role of factors other than eustasy in stratigraphic architecture.

Chapters 2 and 3 differ slightly from the published manuscripts to reflect corrections requested by the Examining Board and requirements of thesis formatting.
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CHAPTER 1: INTRODUCTION

The northern Appalachian Basin, the Ottawa Embayment, and the western Quebec Basin define an arcuate (SW-to-NE) trend of relatively undeformed sedimentary rock of Early Paleozoic age inboard of the Appalachian orogen, crossing the international boundary of the United States and Canada (Fig. 1.1A). This geographic region, and these sedimentary rocks beneath, form the mid-19th century birthplace of sedimentary geological analysis in North America in association with emergence of the New York State Geological Survey (1836) and the Geological Survey of Canada (1842) (Vodden, 1992; Dott, 2005). Over time, the Middle through Upper Ordovician succession of mostly carbonate rock within the more expansive Appalachian and Quebec basins has become testing grounds for application of modern integrated stratigraphy, an approach that achieves improved understanding of continuity of depositional patterns and stratigraphic architecture through integration of lithostratigraphy, biostratigraphy, sequence stratigraphy, chemostratigraphy ($\delta^{13}$C, Sr-isotope), bentonite stratigraphy and geochronology (e.g., Bergström et al. 2010; Sell et al. 2015; Mitchell et al. 2004; Barnes, 1964; Kay, 1937; Holland and Patzkowsky 1996; Joy et al. 2000; Brett et al. 2004; Ettensohn, 2008; Lavoie, 2008). In parallel, improved understanding of development of the Appalachian orogen (Hatcher, 2010; Karabinos et al., 2017) has allowed the beginning of fitting Cambrian and Ordovician platform dynamics to plate-margin evolution (Lavoie, 2008; Ettensohn, 2008; Macdonald et al., 2017). Yet, despite advances over ~150 years of research there remain conflicting interpretations of eustatic versus tectonic controls on sedimentation for the Middle to Upper Ordovician succession of the Appalachian Basin that underlies much of eastern United States (Fig. 1.1B; Joy et al., 2000; Brett et al., 2004; Sell et al., 2015).
Fig. 1.1: (A) General geologic elements and geographic features of the Ottawa Embayment, forms a westward extension of the St. Lawrence Platform. The Ottawa-Bonnechère graben (OBG) fault distribution highlighted (dash lines), illustrates a west-directed curvature (from E-W to NW-SE). The OE is delimited from the adjacent western Quebec Basin by the Oka-Beauharnois anticline (O-B) and from the Appalachian Basin by the Frontenac arch (FA). Modified from Sanford (1993). (B) Paleogeography of the Ottawa Embayment (OE) relative to adjacent basins in the United States and southern Ontario (Appalachian Basin, Michigan Basin, Illinois and Black Warrior Basin), and nearby structural features at ~453 Ma. Abbreviation: AA, Algonquin arch; LA, Laurentian arch; TG, Timiskaming Graben; GF, Grenville Front; AH, Adirondack highland; fa, Findlay arch; CA, Cincinnati arch; AF, Appalachian structural front.
This conflict in interpretation is of particular interest because the Middle to Upper Ordovician succession represents deposition during a period of plate convergence with orogen development along the southern Laurentian plate boundary (Fig. 1.1B). The adjacent basins represent the tectonic foreland (Ettensohn, 2008; Lavoie, 2008), where the record of this convergence is preserved.

Compared to these two much larger bounding basins, a detailed record of Middle to Upper Ordovician foreland stratigraphy and deposition within the Ottawa Embayment is fragmentary at best, but a problem resolved by this study. Previous work has emphasized eustatic (global) control on sedimentation superimposed on a background of regional foreland subsidence (Johnson et al., 1992; Sanford, 1993). The Ottawa Embayment is a structurally defined tectonic sub-province of the Central St. Lawrence Platform (Fig. 1.1; Sanford, 1993; Salad Hersi, 1997). The embayment’s limits, defined by erosional and structural margins, characterizes a WNW-oriented indentation of Paleozoic strata along the southern limit of the Canadian Shield. The orientation parallels an underlying structural axis of an intracratonic Neoproterozoic fault system (Burke and Dewey, 1973; Kumarapeli, 1985; Mc Clausland et al., 2007; Fig. 1.1A), now manifest through crustal reactivation as the Ottawa-Bonnechère graben (Kay, 1942; Kumarapeli and Saull, 1966; Kumarapeli, 1985). Only Lower Paleozoic sedimentary rocks are preserved in the embayment and comprise local expressions of two cratonic megasequences (Sloss 1963; Fritz et al., 2012): the Sauk, and Tippecanoe I (Fig. 1.2). A Cambrian to Early Ordovician succession defines platform-interior deposition along the Laurentian trailing margin facing the developing Iapetus Ocean. A younger Middle to Late Ordovician rock succession representing initial development of a foreland-interior platform distal to
**Fig. 1.2:** Generalized lithostratigraphy of the Lower Paleozoic succession that underlies the Ottawa Embayment, central Canada. The Millbrig Bentonite (M) and other recognized bentonites are indicated (red bars). The term megasequence follows Fritz et al. (2012). Chronostratigraphic nomenclatures are based on Ogg et al. (2008) and geochronology of orogen development along the southern Laurentian margin is based on Macdonald et al. (2017). Abbreviation: E, Eastview Member; NP, Nepean Point Member; Eden, Edenian; Mays, Maysvillian; Rich, Richmondian. Chazyan is only a regionally defined stage, and Cambrian and Lower Ordovician stages are not differentiated. The slant lines represents hiatus.
the developing orogen, then foreland basin development following platform foundering in response to orogen development through the Late Ordovician (Johnson et al., 1992). Including the mentioned fragmentary record of details relating to foreland development in the embayment, there has been a longstanding unresolved lithostratigraphic problem when comparing the embayment’s foreland stratigraphy with the equivalent extrabasinal succession in the northern Appalachian Basin (Kay, 1937). The type lower Trentonian (now lower Chatfieldian) succession for eastern North America was defined from a quarry near Rockland in the eastern Ottawa Valley (Raymond, 1914). It became apparent, however, that sedimentary attributes of this succession could not be correlated beyond the embayment’s limits (Kay, 1937; Barnes, 1968; Cameron and Mangion, 1977) and a new reference section was established within the northern Appalachian Basin west of Kingston (Fig. 1.1; Cameron and Mangion, 1977). Since then, the foreland stratigraphy of the Ottawa Valley has largely been ignored by workers studying the Appalachian Basin and equivalent successions in basins underlying the central United States (Fig. 1.1B). Most importantly, previous interpretations calling for eustatic influence across the southern Laurentian platform must, by definition, have also occurred in to the Ottawa Embayment and left distinct signatures in the sedimentary succession. However, this has gone untested.

There has been varied formation nomenclature applied to the Ottawa Embayment since Raymond’s (1914) time (Johnson et al., 1992; Sanford, 1993). In my study, formal definition and redefinition of some formations are presented (Chapter 2), and use of other formation names are dictated by a combination of their regionally mappable lithic characteristics carried from the northern Appalachian Basin, stratigraphic priority, or
unique presence in the embayment (Wilson, 1946; Salad Hersi and Dix, 1999; Landing and Westrop, 2006; this study). In particular, the term Bobcaygeon Formation defined from equivalent central Ontario stratigraphy and previously applied to the embayment by Liberty (1969) is not used in my study of the Ottawa Embayment. Stratigraphic analysis shows a high-order lithic resolution allowing application of previous lithic terms long used in the basin (e.g., Hull and Rockland formations,) coupled with definition of a new formation, L’Orignal (see Chapter 2).

My study offers new contributions about the role of tectonics in development of the Middle to Upper Ordovician foreland succession of eastern North America. It presents an integrative study of the mostly carbonate rocks underlying the Ottawa Embayment by developing intrabasinal (embayment-wide) frameworks for lithostratigraphy and sedimentology (Chapter 2), $\delta^{13}$C chemostratigraphy (Chapter 3), and sequence stratigraphy (Chapter 4). For each, comparison with coeval successions in basins extending across the eastern and central United States is performed to evaluate inter-regional continuity (Fig. 1.1B). Comparison with the western Quebec Basin is more difficult (but is shown with respect to lithostratigraphy in Chapter 2) due to prominent condensed intervals within the Upper Ordovician interval and uncertainty with basin-to-platform correlations. By defining periods of inter-regional continuity versus discordance in depositional patterns, one can begin to provide the basis to differentiate eustasy, tectonism, and sediment supply as controls on sedimentation.
CHAPTER 2: LITHOSTRATIGRAPHY OF THE UPPER TURINIAN-LOWER CHATFIELDIAN (UPPER ORDOVICIAN) FORELAND SUCCESSION, AND A U-Pb ID-TIMS DATE FOR THE MILLBRIG VOLCANIC ASH BED IN THE OTTAWA EMBAYMENT

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Abstract

Three stages of carbonate-platform development are preserved in the upper Turinian – lower Chatfieldian succession of the Ottawa Group in the Ottawa Embayment and represent deposition along the Late Ordovician Taconic foreland interior of paleo-southern Laurentia. Compared with contemporary stratigraphy in the adjacent northern Appalachian (southern Ontario, New York State) and western Quebec basins, the intermediate Stage 2 succession, which brackets the Turinian–Chatfieldian boundary, preserves embayment-specific stratigraphic patterns. These include: (i) dramatic west-to-east thickening of the upper Turinian Watertown Formation that defines differential subsidence along the present axis of the embayment, (ii) post-Watertown base-level fall defined by appearance of shoreface siliciclastics, (iii) early Chatfieldian marine transgression represented by the proposed L’Orignal Formation that is coeval with but lithologically distinct from the Selby Formation in the northern Appalachian Basin, and (iv) platform segmentation that resulted in a depositional mosaic of shallow banks (Rockland Formation) and equivalent deeper water micro-seaways (lower Hull Formation). The latter event immediately follows accumulation of the Millbrig bentonite, here dated at 453.36 ± 0.38 Ma. Bracketing this interval that contains these local stratigraphic patterns are the bounding stages (1 and 3) represented by the upper Turinian Lowville Formation and middle Chatfieldian Hull Formation, respectively. These intervals contain facies attributes in common with the adjacent basins, and interpreted to have been deposited during first warm, then cooler inter-regional oceanographic conditions. Stage 2 identifies a structurally controlled transition between these end-member stages; a far-field response in the foreland interior, localized along the axis of a
late Precambrian fault system, to contemporary change in subsidence rates and tectonomagmatic events along the Laurentian margin.

**Introduction**

Upper Ordovician (Turinian–Chatfieldian) strata of eastern North America include the classic Black River and Trenton group succession of New York that has received considerable attention since the mid-1800s (Fisher 1962; Cameron and Mangion 1977; Brett and Baird 2002, and references therein). Over the last two decades, improved resolution of local and regional depositional architectures through the eastern and central United States has emerged through integration of bentonite stratigraphy and geochronology, sequence stratigraphy, and δ¹³C profiles (Holland and Patzkowsky 1996, 1998; Kolata et al. 1996; Joy et al. 2000; Mitchell et al. 2004; Brett et al. 2004; Bergström et al. 2010; Sell et al. 2015). This provides an improved basis from which to differentiate local (sedimentary, tectonic) versus more expansive (tectonic, eustatic) controls on base level along the ancient Laurentian margin and timing with structural development of the distal orogen (Karabinos et al. 2017; Macdonald et al. 2017). In particular, this work supports the transformation from warm to cooler water carbonate deposition across the Laurentian margin through the early Chatfieldian (e.g., Keith 1988; Brookfield 1988; Patzkowsky and Holland 1993; Lavoie 1995; Pope and Read 1997; Holland and Patzkowsky 1998) driven by increased subsidence rates driven by foreland basin tectonics (Ettensohn 2008). In addition, the Millbrig bentonite (the platform interior record of voluminous distal outpouring of volcanic ejecta (Huff et al. 1996) defines an
**Fig. 2.1:** Geological elements of the Ottawa Embayment (OE) and study localities. The embayment is bounded by Precambrian continental crust and is separated from the western Quebec Basin (nQB) by the Oka–Beauharnois anticline, and northern limits of the Appalachian Basin (nAB) by the Frontenac Arch (FtA). Section locations are indicated (see Appendix A for coordinates). Other regional structures include the Adirondack and Laurentian highlands; the Grenville mafic dyke swarm (green); and Precambrian lithotectonic terrains: CMB, Central Metasedimentary Belt; FA, Frontenac–Adirondack lowland terrain; and MAH, Morin–Adirondack highland terrain. Possible subsurface extension and isolated blocks of the MAH terrain beneath the embayment are indicated by dashed grey lines. Paleozoic megasequences are divided into Sauk and (following Dix, 2012) Tippecanoe I Megasquence, divided into the initial foreland platform and subsequent foreland basin-fill succession. Fault pattern defines a prominent curvature of the Ottawa–Bonnechere graben extending northwest of the embayment, and the Gloucester and Rideau faults are highlighted. Map is based on Baer et al. (1977), Gupta (1991), Sanford and Arnott (2010), and Bleeker et al. (2011).--
important stratigraphic baseline for correlation, the base of the Chatfieldian Stage (Leslie and Bergström 1995).

The Ottawa Embayment, positioned between the northern Appalachian Basin and Quebec Basin (Fig. 2.1), characterizes a significant information gap with respect to this modern database related to lower Chatfieldian stratigraphy. As the birthplace of lower Trenton stratigraphy (Raymond 1914), the Rockland and overlying Hull formations (Fig. 2) formed the basis of initial regional correlation through eastern North America (Kay 1933, 1937), and the Rocklandian remains a regional stage or substage (Fig. 2.2; Webby et al. 2004). Kay (1937) erected two successive reference sections, now the Selby and Napanee formations (Kay 1968a; Cameron and Mangion 1977), in the northern Appalachian Basin (Ontario) as an age-equivalent succession to the Rockland Formation (Fig. 2.2). These are now part of the standard Mohawkian lithostratigraphic section (Cameron and Mangion 1977), yet formation lithologies are dissimilar to the type Rockland (Barnes, 1968). Use of this extrabasinal formation nomenclature in the Ottawa Embayment (Fig. 2.2; e.g., Kay 1968b; Cameron and Mangion 1977; Barta et al. 2007) tacitly assumes common interbasinal depositional attributes and base-level controls.

We present a new lithostratigraphic framework for the upper Turinian–lower Chatfieldian succession in the Ottawa Embayment (Fig. 2.2), integrated with a U–Pb isotope dilution–thermal ionization mass spectrometry (ID–TIMS) date for the Millbrig bentonite. The purpose of this study is to better understand the origins of lithic dissimilarity between the Rockland Formation in the Ottawa Embayment and the extrabasinal Selby–Napanee succession in the northern Appalachian Basin. From this emerges a better understanding of local versus regional controls on depositional systems and stratigraphy in
Fig. 2.2: Selected stratigraphic nomenclature associated with upper Turinian – lower Chatfieldian strata in the Ottawa Embayment, southern Ontario and New York State (after Mitchell et al., 2004). Interpreted positions of the Blackriveran-Trentonian and equivalent stratigraphic boundary (thick black lines), and the Millbrig bentonite and Turinian-Chatfieldian stage boundary (red lines) are indicated. Different positions (a, b) have been interpreted (Brett et al., 2004; Sell et al. 2015). Abbreviations: Laur, Laurentian; Mbr, Member. Global chronostratigraphy is based on Ogg et al. (2008), and use of the term megasequence follows Fritz et al. (2012).
### northern Appalachian Basin

<table>
<thead>
<tr>
<th>Global Series/Stage</th>
<th>Laurentian Series/Stage</th>
<th>New York</th>
<th>Kingston - Napanee</th>
<th>Ottawa Embayment</th>
<th>Quebec Basin</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Ordovician</td>
<td>Katian</td>
<td>Hull</td>
<td>Kings Fall</td>
<td>Hull</td>
<td>H1</td>
</tr>
<tr>
<td></td>
<td>Chertfeldian</td>
<td></td>
<td></td>
<td></td>
<td>H2</td>
</tr>
<tr>
<td></td>
<td>Rockland</td>
<td></td>
<td></td>
<td></td>
<td>H3</td>
</tr>
<tr>
<td></td>
<td>Flandrian</td>
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</tr>
<tr>
<td></td>
<td>Blackriveran</td>
<td></td>
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<tr>
<td></td>
<td>Tippecanoe I Megasquence (Part)</td>
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<tr>
<td></td>
<td>Glenburnie</td>
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<tr>
<td></td>
<td>Leray</td>
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<tr>
<td></td>
<td>Lowville</td>
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<td></td>
<td>Pamela</td>
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<td></td>
<td>Upper</td>
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<td>Lower</td>
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<td></td>
<td>Kay 1937</td>
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<tr>
<td></td>
<td>a) Brett et al., 2004</td>
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<tr>
<td></td>
<td>b) Self et al. 2015</td>
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</tr>
</tbody>
</table>

### Ottawa Embayment

<table>
<thead>
<tr>
<th>Scottish Series/Stage</th>
<th>Lower</th>
<th>Rockland</th>
<th>Upper</th>
</tr>
</thead>
<tbody>
<tr>
<td>L'Original</td>
<td></td>
<td>Rockland</td>
<td></td>
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</tbody>
</table>

### Quebec Basin

<table>
<thead>
<tr>
<th>Scottish Series/Stage</th>
<th>Lower</th>
<th>Rockland</th>
<th>Upper</th>
</tr>
</thead>
<tbody>
<tr>
<td>L'Original</td>
<td></td>
<td>Rockland</td>
<td></td>
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</tbody>
</table>

- H1, H2, H3, H4: Different layers or formations in the geological strata.
- "upper" and "lower": Refers to different geological eras or periods.
the Ottawa Embayment during a period of substantial oceanographic and tectonic transformations along paleo-southern Laurentia (Ettensohn 2008; Macdonald et al., 2017).

**Geologic Setting**

The Ottawa Embayment forms a geographic indentation of lower Paleozoic strata along the southeastern Canadian Shield (Fig. 2.1; Sanford, 1993). It lies inboard of the Appalachian structural front and is separated from the western Quebec Basin by the Oka–Beauharnois anticline (Fig. 2.1). A narrow corridor of high-angle faults within and extending northwest of the embayment defines the Ottawa–Bonnechère graben (Fig. 2.1; Kay, 1942). This intracratonic fault system lies sub-parallel to the late Neoproterozoic Grenville mafic dyke swarm (590 Ma; Kamo et al. 1995) and cross-cuts, at a high angle, Grenville lithotectonic boundaries of the underlying metamorphic continental crust (Fig. 2.1). It represents a long-lived corridor of structural and thermal events, including: a late Precambrian aborted rift; Early Paleozoic, Mesozoic, and Quaternary faulting; fault reactivation and post-Paleozoic uplift; and sustained seismicity especially along the graben’s northern limit (Burke and Dewey, 1973; Crough, 1981; Roden-Tice et al., 2005, 2012; Rimando and Benn, 2005; Dix and Al Rodhan, 2006; Ma and Eaton, 2007; Dix and Jolicoeur, 2011; McCausland et al., 2007; Nurkhanuly and Dix, 2014; Hardie et al., 2017; Lowe et al., 2018).

A Lower Cambrian through Middle Ordovician stratal succession is present within the Ottawa Embayment (Sanford, 1993; Lowe et al., 2017). It forms the local expression of the cratonic Sauk megasequence (Fritz et al., 2012), and contains platform-interior facies
deposited in a distal location relative to the Laurentian trailing margin (Lavoie et al., 2012). An overlying Middle through Upper Ordovician (Richmonedian) siliciclastic and carbonate succession forms the local expression of the Tippecanoe I megasequence (Fritz et al. 2012). This stratigraphy documents development then collapse and burial of a foreland-interior carbonate platform (Sanford, 1993; Lavoie, 1994).

**Stratigraphic Framework, Ottawa Embayment**

The Turinian-Chatfieldian succession was initially differentiated as a succession of “beds” with different macrofossil assemblages (Raymond, 1914; Wilson, 1936, 1946) that were subsequently raised to formation status as part of the Ottawa Group (Fig. 2.2; Uyeno, 1974). There has been no more recent characterization of macrofossil zonation for the embayment, and summaries of conodont biozonation have been associated with individual sites (Schopf, 1966; Salad Hersi and Dix, 1999), or specific stratigraphic intervals (Barnes, 1964, 1967; Uyeno, 1974).

Stewart Quarry (Loc. 3, Fig. 2.1) exposes the upper Pamelia through Hull formations of the Ottawa Group (Fig. 2.3; Wilson 1921; Salad Hersi and Dix 1999) and has received the greatest attention being host to the type Rockland Formation (Raymond 1914). Wilson’s (1921) Pamelia, Lowville, and Leray beds (Fig. 2.3) appear to be contemporaneous with the respective Blackriveran Pamelia, Lowville, and Chaumont or Watertown succession in the northern Appalachian Basin (Fig. 2.2; Cushing et al., 1910; Kay, 1942; Barnes, 1967; Salad Hersi and Dix, 1999). The lowermost Rockland Formation contains macro- and microfaunal assemblages of mixed Blackriveran and Rocklandian affinities (Wilson, 1921; Barnes, 1967), and the conodont *Belodina*
Fig. 2.3: Summary of previously defined stratigraphic frameworks for Stewart Quarry (Loc. 3, Fig. 2.1), the type section of the Rockland Formation (Raymond, 1914). The Millbrig bentonite has not been found in this section.
*confluens* biozone appears in calcarenite of the Hull Formation that disconformably overlies the Rockland succession (Barta et al., 2007). The disconformity truncates the upper part of the Guttenberg isotope carbon excursion (Fig. 2.3; Barta et al., 2007).

**Methods**

Lithostratigraphy and facies successions are based on road outcrops, core, and quarry sections (Fig. 2.1, Appendix A). Skeletal components and other depositional attributes noted in hand specimens were confirmed from thin-section and binocular microscopy.

Additional methods were used to characterize an interpreted altered volcanic-ash fall deposit at Loc. 1. Grain-size range was determined by laser diffraction using a Beckman Coulter LS 13 320 (Carleton University), with data processed using the program Gradistat (Blott and Pye, 2001). Mineralogy of the <2 µm fraction was determined by X-ray diffractometry (Geological Survey of Canada) using air-dried, glycolated, and heated (550 °C) sections (Appendix B). Biotite grains were handpicked from the wet-sieved sand fraction, and zones of alteration and relict mineralogy were differentiated in polished grain mounts through backscatter electron imagery. Elemental geochemistry of relict biotite was carried out using a Cameca Camebax MBX electron microprobe (Carleton University), with sampling based on rastered (8 × 8 µm) areas, and using a standard silicate package for data collection and reduction. Grain-size separation recovered an abundance of zircon crystals. Several dozen of these were selected for chemical abrasion from which six single grains were selected for U–Pb analysis by ID–TIMS at the Jack Satterly Geochronology Laboratory, University of Toronto (Appendix B).
Upper Turinian - lower Chatfieldian lithostratigraphy, Ottawa Embayment

Lithostratigraphic correlation is shown for a transect (~200 km) oriented parallel to the axis of the embayment (Fig. 2.4). Two sections (Locs. 4 and 5) are more centrally positioned (Fig. 2.4). Lithofacies successions and attributes (Table 2.1) are summarized below.

Lowville Formation

*Lithofacies succession*

The formation disconformably overlies dolostone and, locally, sandstone of the Blackriveran Pamelia Formation (Salad Hersi and Dix 1999; this study). Two facies successions comprise the unit (Table 2.1, Fig. 2.5): a lower (L1) division includes skeletal- and ooid-rich coarse-textured limestones (Fig. 2.6a; Table 2.1; Salad Hersi and Dix 1999), and an upper (L2) division consists of thinly bedded skeletal-bearing mudstone to *Tetradium* rudstone (with fragments oriented prone to bedding) host to locally abundant *Tetradium* mounds (Figs. 2.6a and 2.6b). In the western embayment (Loc. 6 and 9), the mounds consist of skeletal material engulfed in mudstone (Figs. 2.6b and 2.6c). There is pronounced differential compaction beneath the larger mounds, but others appear to fill narrow channels. To the east, *Tetradium* colonies lack the enclosing lime mudstone and are associated with gravel size skeletal limestone, often with a profusion of fragmented *Tetradium* (Wilson, 1946; Salad Hersi and Dix, 1999). Upward expansion of some mounds in the upper <0.60 m of the formation develops discontinuous (over metres) beds of burrowed lime mudstone with scattered *Tetradium* colonies (Fig.
Fig. 2.4: Lithostratigraphic correlation of the Lowville through Hull succession in the Ottawa Embayment along a ~200 km transect extending from near L’Orignal (Loc. 1) to near Arnprior (Loc. 9). All sections are measured from base of exposure with the exception of core comprising Loc. 5. Interpreted bentonites (b) and Millbrig bentonite (M) are indicated. The interpreted isolated character of Rockland strata and lateral equivalence with unit H1 (Hull Formation) are indicated.
<table>
<thead>
<tr>
<th>Formation</th>
<th>Boundary* Type</th>
<th>Division</th>
<th>Thickness (metres)</th>
<th>Lithofacies Associations**</th>
<th>Interpreted Paleoenvironment</th>
<th>Sources***</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lowville</td>
<td>base: 2</td>
<td>L1</td>
<td>4 - 21</td>
<td>skeletal/ooid packstone/grainstone; skeletal mudstone; calcareous shale</td>
<td>peritidal/subtidal, low to high energy</td>
<td>a-c, m</td>
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<td></td>
<td>top: 3</td>
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<td></td>
<td>L2</td>
<td>1 - 7</td>
<td></td>
<td>mudstone (minor packstone and grainstone) with <em>Tetradium</em> mounds, local biostromes, and rudstone</td>
<td>protected, open shelf</td>
<td>a-c, m</td>
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<td></td>
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<td></td>
<td></td>
<td><em>FT</em>: b, br, c, co, g, n, t</td>
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<tr>
<td>Watertown</td>
<td>base: 3</td>
<td>W1</td>
<td>0.4 – 1.8</td>
<td>coarse-grained skeletal packstone, grainstone, and rudstone</td>
<td>high-energy, normal marine</td>
<td>d, m</td>
</tr>
<tr>
<td></td>
<td>top: 4</td>
<td></td>
<td></td>
<td>FT: br, bry, c, co, g, n, t, s</td>
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</tr>
<tr>
<td></td>
<td>W2</td>
<td>2 – 25</td>
<td></td>
<td>(a) thick beds, skeletal and peloidal wackstone; burrowed; locally nodular fabric; beds are stylobound; FT: b, br, c, co, g, n, t</td>
<td>(a) low energy, normal marine, open platform;</td>
<td>c-g, m</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(b) hardground (planar, undulating)</td>
<td>(b) wave erosion</td>
<td>b, m</td>
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<td></td>
<td></td>
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<td>FT: bo (local); c (in depressions)</td>
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<tr>
<td>L’Orignal</td>
<td>base: 3</td>
<td>LO1</td>
<td>trace</td>
<td>admixed bentonite and crinoid debris, quartz arenite with rare “blackened” phosphatic lithoclasts (quartz arenite, wacke)</td>
<td>reworked volcanic ash shallow shoreface, with near-surface reworking and phosphatization</td>
<td>h, i, m</td>
</tr>
<tr>
<td></td>
<td>top: 3</td>
<td>LO2</td>
<td>&lt; .10</td>
<td>skeletal packstone, and skeletal grainstone</td>
<td>normal-marine transgression</td>
<td>b, c, m</td>
</tr>
<tr>
<td></td>
<td>LO3a</td>
<td></td>
<td>&lt; 0.5</td>
<td>FT: a (Vermiporella sp.), b, br, c, bry (<em>Heterotrema</em> sp. <em>Stictopora fenestrata</em>), g, o, m (<em>Girvanella</em>), sp, t</td>
<td>protected subtidal;</td>
<td>b, m</td>
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<tr>
<td></td>
<td>LO3b</td>
<td></td>
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<tr>
<td></td>
<td>LO4</td>
<td></td>
<td>0.5 to 9</td>
<td>rhythmic bedding of skeletal</td>
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</tr>
<tr>
<td>Location</td>
<td>Base</td>
<td>Top</td>
<td>Description</td>
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<tr>
<td>LO5</td>
<td>0.12</td>
<td></td>
<td>packstone (or mudstone: Loc. 4) and calcareous shale. FT: as above; but, at Loc. 4; c, bu shale within which is a 5-cm-thick bentonite. volcanic ash deposition with shale transgression mostly a low-energy bank with repeated shut-down of carbonate production. b, m</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rockland (revised)</td>
<td>5</td>
<td>3</td>
<td>Rhythmic interbeds: fossiliferous lime mudstone and calcareous shale rare skeletal grainstone/packstone. FT: br, c, g, co (rugose). (a) normal marine; deeper platform or bank-margin slope or channel-fill. j, m</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hull (revised)</td>
<td>3</td>
<td>1</td>
<td>(a) thin interbeds of skeletal-rich rudstone/grainstone and skeletal-rich shale; at Loc. 1: large scale cross-beds and primary dip (SSW: 10-20°) FT: a (Solenopora sp.), br, br, bry (cryptostomid, treptostomid), c, n, t. (b) at Loc. 4: 3-m-thick succession of interbedded skeletal mudstone and shale bounded by (a) FT: as above. (b) rhythmic pattern of higher and lower energy related to sedimentation. j, m</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H2</td>
<td>2</td>
<td>26</td>
<td>Lozenge-shaped fine-grained crinoidal grainstone, separated, by thin beds laminae, and partings of shale. volcanic ash deposition with shale transgression. Low-energy flank, open platform shoal. k, l, m</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>H3</td>
<td>4-13</td>
<td></td>
<td>(a) skeletal-rich (mostly crinoid) grainstone/rudstone; amalgamated bedding, tabular cross-bedded (locally reversed), erosional surfaces; (a) high-energy shoal; j, k, l, m (b) protected intershoal or landward facies. m</td>
<td></td>
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</tr>
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local disrupted, convoluted beds
FT: br, c, co (Tetradium, other colonial forms)
(b) at Loc. 9: interbedded lime mudstone and shale; bioturbated skeletal packstone
FT: as above

* 1, conformable; 2, abrupt (does not include evidence of erosion or stylolites); 3, abrupt, but locally erosional; 4, erosional (regional); 5, stylolitic
** FT = fossil types: a, calcareous algae (type defined); b, bivalve; bo, borings (Trypanites); bu, burrows; br, brachiopod; bry, bryozoan; c, crinoid; co, coral (type defined); g, gastropod; m, microbial (type defined) n, nautaloid; o, ostracodes; s, stromatoporoids; sp, sponge spicules; t, trilobite
*** a, Salad Hersi and Dix, 1999; b, Flügel (2010); c, James and Jones, 2016; d, Kay, 1937, 1942; e, Wilson, 1946; f, Barnes, 1967; g, Cameron and Mangion, 1977; h, Birch, 1980; i, Pufahl, 2010; j, Titus and Cameron, 1976; k, Uyeno, 1974; l, Mehrtens, 1988b; m, this study
2.6a and 2.6d). The formation top is most often abrupt and planar, but an erosional surface is demonstrated at Locs. 3 and 10 by truncation of underlying mound-bearing facies (Figs. 2.6c and 2.6d). At Stewart Quarry (Loc. 3), this includes truncation of shallowly dipping *Tetradium*-bearing rudstone beds that appear to flank a mud-rich *Tetradium* mound (Fig. 2.6c).

The Lowville Formation has a relatively uniform regional thickness, 8–9 m, except proximal to two regional fault traces where thicknesses reach up to 25 m (Fig. 2.7a). Regionally, the upper (L2) division remains <2 m thick across the study area.

*Environmental Interpretation*

The lower division facies succession is interpreted to characterize a regional paleogeography of high-energy shallow subtidal shoals and low-energy muddy lagoons (Salad Hersi and Dix, 1999). We interpret the upper division to characterize initial deepening, producing (in the western embayment) quieter shallow-subtidal platform conditions. Very coarse-grained fragmental limestone, especially to the east, suggests local reworking of *Tetradium* thickets due to storm activity. The upward transition from *Tetradium* mounds to single beds may define a response to shallowing, and formation of single-bed biostromes (cf. Kershaw, 1994).

*Watertown Formation*

*Lithofacies succession*

The formation is divided vertically into two facies successions (Fig. 2.5). A lower (W1) division consists of thinly bedded (<15 cm) coarse-grained skeletal packstone, grainstone,
Fig. 2.5: Stratigraphic motifs (without scale) of lithofacies and facies associations within the Lowville through Hull succession of the Ottawa Embayment. See Table 2.1 for details. Abbreviation: hg, hardground.
and rudstone equivalents (Table 2.1; Fig. 2.6e). It appears at Loc. 3, and thickens westward to ~1.8 m at Loc. 9. The upper (W2) division is represented by thick to very thick (30–190 cm), stylolite-bound beds of burrowed skeletal and peloidal wackestone (Fig. 2.6e; Table 2.1; see also Barnes 1967; Salad Hersi and Dix 1999). Rare skeletal grainstones appear at some localities. The beds are host to numerous planar to gently undulating hardgrounds, some with local borings, and often erosionally truncating subvertical and 3-dimensional burrows networks (Fig. 2.6f). Local crinoid debris differentially fills erosional paleodepressions and intersected burrow networks along a given hardground surface (Fig. 2.6f). Laterally amalgamated hardgrounds (Fig. 2.6f) identify a multigenerational history for some single surfaces. The W1 and W2 divisions both contain diverse fossil assemblages of whole to fragmented fossils (Table 2.1). A biohorizon of bulbous stromatoporoids occurs within 30 cm above the W2 division base.

The formation varies greatly (<4–25 m) in thickness across the embayment. Sites of greater thickness define a narrow corridor parallel to the embayment’s axis (Fig. 2.7b). Regional westward thickening of the W1 division coincides with westward thinning of the formation to <4 m at Locs. 2, 8, 6, and 9 (Fig. 2.7b), sites at which closely spaced hardgrounds are abundant.

*Environmental Interpretation*

The coarse-textured nature and westward thickening of the lower (W1) division is interpreted as a high-energy transgressive deposit developed above a marine ravinement surface during west-directed translation of the shoreline (Catuneanu et al. 2011) that
Fig. 2.6: Lithic characteristics of the Lowville and Watertown formations, Ottawa Embayment. (a) Lower (L) and upper (U) divisions (boundary highlighted in black) of the Lowville Formation at Loc. 9 with upward coalescence of two adjacent large Tetradium mounds (black arrows) into a laterally extensive muddy biostrome (white arrow). Hammer for scale. (b) Close-up of a Tetradium mound with skeletal elements engulfed by mudstone. Coin (Canadian quarter, 23.81 mm diameter) for scale. (c) Disconformable contact (arrow) between the Lowville and lower Watertown (W) formations at Loc. 3. The upper metre of the Lowville Formation shows an interpreted biostrome with muddy core (c) adjacent flanking beds (f) with abundant Tetradium fragments. Backpack for scale. (d) A Tetradium biostromal unit (b), overlying thinly bedded mudstone (m) of the upper division (L1) of the Lowville Formation, is truncated by a planar disconformity (arrows) and overlain by wackestone (w) of the Watertown Formation. Loc. 10 (for location, see Appendix A). (e) Basal coarse-grained lithofacies (W1) of the Watertown Formation beneath the more thick-bedded muddy lithofacies (W2). Loc. 9, hammer for scale. (f) Lateral coalescence of two hardgrounds (arrows) within lithofacies W2 of the Watertown Formation. Loc. 9, coin (quarter, 23.81 mm diameter) for scale. (g) Undulating erosional top (white arrow) of the Watertown Formation (w) overlain by thinly bedded limestone of the L’Orignal Formation (p) at Loc. 9. This exposure highlights a differentially truncated burrow network in Watertown strata partially filled with whitish crinoid-bearing clayey lime mudstone interpreted as a reworked bentonite overlying the erode Watertown strata. Marker for scale.
extends westward across the Lowville paleoplatform. Appearance of the W2 facies association identifies subsequent deepening with continued transgression, muddier sediment accumulating in a lower energy, subtidal, normal-marine setting (Barnes 1967; Cameron 1968). The W1 to W2 stratigraphic motif (Fig. 2.5) is similar to a style of carbonate platform succession (James and Jones 2016) wherein hardground development is promoted during periods of temporary lowering of sea level resulting in increased seafloor scouring, impeded sediment accumulation, and warmer temperatures (McFarlane 1992; Nelson and James 2000). Repeated, this effect created net stratigraphic condensation of the Watertown Formation in the western embayment, whereas much thicker successions are preserved along the embayment axis to the east. Local stratigraphic condensation may explain anomalous local thinning of the formation along the northern limit of the embayment (Locs. 2, 5, and 8; Fig. 2.7b).

**L’Orignal Formation (proposed)**

The formation stratotype (Loc. 1; Fig. 2.8a) is exposed in an active quarry near L’Orignal, Ontario. The unit disconformably overlies the Watertown Formation, and is overlain locally by two formations: the revised Hull Formation (Locs. 1, 4, 5, 6, and 9; Figs. 2.4, 2.8b, and 2.8c) and the revised Rockland Formation (Loc. 2 and 3; Fig. 4). At Loc. 3, the L’Orignal Formation is equivalent to the lower type Rockland Formation. The L’Orignal Formation exhibits a uniform regional thickness of <2 m, but with anomalous thickening (~9 m) at Loc. 5 (Fig. 2.7c). General formation characteristics (Table 2.2) and stratotype description (Appendix C) are expanded upon below.
**Fig. 2.7:** Geographic patterns of formation thickness in the Ottawa Embayment. Contours are in metres, with different intervals according to formation thickness. Sites 7 and 8 are illustrated in Appendix D. Circle colours: purple, complete formation; grey, not exposed; open, incomplete due to present-day erosional surface. For the Lowville Formation, additional sites (green boxes) are from Salad Hersi (1997—see his Fig. 1.1 and 3.6, and his Appendix: a, Loc. 10; b, Loc. 11; c, Loc. 12; d, Loc. 15; e, Loc. 27; f, Loc. 34; g, Loc. 45; h, Loc. 47; i, Loc. 48).
Lithofacies Succession

Basal facies (LO1–3a) of the formation vary locally (Fig. 2.5, Table 2.1). The oldest facies (LO1, Table 2.1) is known from Loc. 9 only and represented by erosional remnants of a whitish indurated clay to claystone with pockets of crinoidal debris. This facies partially fills and mantles erosionally truncated burrow networks exposed along the post-Watertown disconformity (Fig. 2.6g). Beneath the disconformity, the normally grey limestone appears differentially bleached below occurrences of LO1. A second basal facies (LO2) is represented at Loc. 4 by a thin (<5 cm) bed of fine-grained horizontally burrowed quartz arenite. It contains outsized gravel-sized quartz grains, brachiopod (lingulid) shell fragments, rare lithoclasts of phosphatic fine- to coarse-grained arenite and wacke, and phosphatic claystone. Lithoclasts are blackened due to manganese oxide (Fig. 2.8d), and local pyrite (framboid)-bearing clasts are encircled by broken rims of iron oxide. Evidence of a more widespread distribution of this basal facies is the presence of lithoclasts of similar quartz arenite elsewhere in the basal carbonate facies. The third local basal facies (LO3a) of the formation forms the lower 45 cm of the formation at Loc. 3, and consists of skeletal grainstone intercalated with laminae of shale. There are local ball-and-pillow structures illustrating soft-sediment deformation and local amalgamated grainstone beds. Above this varied basal distribution of facies, the remainder of the formation consists of thinly bedded fossiliferous packstone and wackestone (facies LO3b, Fig. 2.5) rhythmically intercalated with thin beds, laminae, and partings of shale (LO5; Fig. 2.8a). An exception occurs at Loc. 6, where most of the carbonate fraction consists of burrowed skeletal mudstone (facies LO4, Fig. 2.5).
**Fig. 2.8:** Outcrop and petrography characteristics of the L’Orignal Formation. (a) Typical stratigraphic fabric of the formation (Loc. 1, type section) consisting of thinly interbedded skeletal packstone and shale (between arrows), with a prominent recessive shale (upper arrow) containing the Millbrig bentonite. Hammer for scale. (b) Erosional truncation of burrowed lime mudstone (arrow and highlighted in black) forming the boundary between the top of the L’Orignal Formation (L) and grainstone of Unit H1 of the Hull Formation (H) at Loc. 9. Scale card is 8 cm wide. (c) Exposure of the L’Orignal Formation (between the two arrows) at Loc. 1 overlying the Watertown Formation (a), and the prominent greyish Unit H1 of the Hull Formation (c). Thickness of the Hull Formation to the quarry bench is 5m. (d) Thin section microphotograph of basal quartz arenite (Loc. 4) showing bimodal distribution of quartz with iron oxide rims, phosphatic brachiopod shell (p) fragment, and MnO-altered phosphatic clast (m). (e) Colourless and clear zircons crystals recovered from the bentonite at Loc. 1. Scale bar 200µ. (f) Electron backscatter image of a biotite crystal from the bentonite at Loc. 1, with narrow relict zones of biotite (bt) within extensive chlorite (ch) and illite (i) alteration. Scale bar = 60µ.
Table 2.2. Definition and attributes of the proposed L’Orignal Formation

**Formation Stratotype**
The formation is named for the exposure in an active quarry near L’Orignal, Ontario (Loc. 1, Fig. 2.1): 45°35.727’ N, 74°45.533’ W; NTS 31 C/05, UTM zone 18; and 518550 (easting) and 5039050 (northing). Locs. 2 - 6 (Fig. 2.1; Appendix A) are reference sections.

**Stratigraphic Boundaries**
The formation disconformably overlies the Watertown Formation at all localities, and is overlain disconformably by the Hull Formation at Locs. 1, 4, 5, 6, and 9; and is overlain with stylolitic contact by the revised Rockland Formation at Locs. 2 and 3. At Loc. 3, the formation forms the lower ~1.5 metres of the original Rockland Formation (Wilson, 1921).

**Lithologic Character**
Most of the formation consists of thinly bedded fossiliferous muddy limestone (packstone, local grainstone) intercalated with very thin shale beds, laminae, and partings. The basal (< 10 cm) lithofacies varies locally: at Loc. 9, limestone overlies crinoidal-bearing whitish clay (interpreted as altered volcanic ash); at Loc. 4, lingulid-bearing phosphatic quartz arenite; at remaining localities, rounded quartz arenite lithoclasts occur in packstone.

**Characteristic Fossils**
There is an abundance of fossils (Table 2.1): crinoids, bivalves, brachiopods, trilobites, calcareous algae (*Vermiporella* sp.).

**Thickness**
Regionally, thickness is about 1-2 m, with anomalous thickening (~9 m) at Loc. 5.

**Interpreted Depositional Environment**
Varied basal transgressive succession succeeded by quiet shallow subtidal, yet marked by periods of siliciclastic input that temporarily smothered carbonate accumulation.

**Geologic Age**
The unit is of early Rocklandian age: at Loc. 3, the formation contains “mixed” Blackriveran-Rocklandian macrofaunal and microfaunal assemblages (Wilson, 1921; Barnes, 1964, 1967), and contains the first appearance of the brachiopod *Triplesia cuspidata* that is an indicator of lower Trenton strata (Kay, 1937, 1942; Cameron and Mangion, 1977). The top of the stratotype includes the interpreted Millbrig bentonite, dated at 453.36 ± 0.38 Ma (this study).

**Correlation**
The lithostratigraphic position identifies a stratigraphic equivalence to the Selby Formation of southern Ontario and northern New York State.
At Loc. 1, the upper 10 cm of the formation forms a prominent recessive interval consisting mostly of shale, but with a 5-cm-thick whitish clay bed (see Appendix C for stratigraphic details) bearing abundant sand-size lustrous mica grains and, with further grain segregation, an abundance of euhedral zircons (Fig. 2.8e). About 95% of the clay bed is composed of illite, mixed layer illite/montmorillonite, smectite, kaolinite, and biotite (Fig. 2.9a). The silt-to-sand size fraction displays a bimodal distribution (Fig. 2.9b) of biotite, zircon, and crystalline aggregations of pyrite. Biotite grain size decreases upward through the clay bed coincident with an increased mixture of grey silt. Mica grains contain relict zones of biotite within chlorite and illite alteration (Fig. 2.8f). Biotite composition is compatible with a calc-alkaline subduction source (Fig. 2.9c; Abdel-Rahman 1994).

Depositional Environment

Basal facies of the L’Orignal Formation identify three distinct depositional systems developed across the Watertown paleoplatform. The white clay (Fig. 2.6g) is texturally similar to other thin beds found throughout the Upper Ordovician succession of eastern North America and is interpreted as the distal record volcanic eruptions (Kay, 1935; Sell et al., 2015). Bleached limestone beneath pockets of this clay represents an alteration product arising from pore-fluids passing through these naturally active bleaching clays (Nutting, 1933). The locally preserved basal quartz arenite (Fig. 2.8d), and lithoclasts of similar composition in lowermost carbonate facies elsewhere, identify a once more widespread siliciclastic facies that postdates the altered ash-fall deposit. During the Ordovician, lingulid (inarticulate) brachiopods occupied shallow to deep-water settings (Harper et al. 2004). As the quartz arenite at Loc. 4 is bounded stratigraphically by
Fig. 2.9: Grain-size distribution, mineralogy, provenance, and the age of the bentonite capping the L’Orignal Formation at Loc. 1. (a) Superimposed diffractograms (relative to d-spacing) of air-derived, glycolated, and heated runs of the clay-size fraction. Abbreviations: i, illite; m, montmorillonite; i/m, mixed layer; s, smectite; b, biotite; k, kaolinite. Montmorillonite is separately defined from mixed smectite-illite layers. (b) Bimodal distribution of fine- and medium-grained sand fractions consisting of biotite, zircon, and aggregates of pyrite. The clay-size mode is likely an artefact related to residual clay retained during washing. (c) Oxide discriminant diagram (Abdel-Rahman 1994) suggests a mostly calc-alkaline (subduction) source for relict biotite. (d) Age summary of the bentonite in the uppermost L’Orignal Formation: (left) U–Pb Concordia diagram showing isotope dilution – thermal ionization mass spectrometry results for six, single, chemically abraded zircon crystals. The six overlapping 206Pb/238U dates give a weighted mean age of 453.36 ± 0.38 Ma (2σ; MSWD = 0.52), which is interpreted to approximate the time of deposition of the bentonite. See Appendix E for accompanying data table (right) Plot of the six (Z1–Z6) 206Pb/238U ages with weighted mean and 2-σ age error represented in the grey horizontal rectangle.
normal-marine shallow-water facies, we interpret the brachiopods and associated horizontal burrows as part of a middle to lower shoreface setting (Howard and Frey 1983; Plint 2010). Local phosphatization in such a setting (e.g., Type A1 of Birch 1980) may arise from phosphate-rich waters carried into the platform interior from more seaward upwelling (Pufahl 2010). But an alternate origin with relevance to this post-Watertown stratigraphy is enhanced stimulation of biological productivity during marine reworking of volcanic ash (Felitsyn and Kirianov 2002). The presence of both pyritic- and MnO-bearing lithoclasts illustrate differential oxic and sulphidic diagenesis (Price 1976), as often related to cycles of near-surface burial, then exhumation leading to intensification of marine phosphatization (Pufahl 2010).

The third post-Watertown depositional system is multifold producing the regional mosaic of initial high-energy (grainstone) to low-energy (mudstone) lowermost carbonate facies of the formation. With continued transgression, a more uniform deeper setting of moderate- to low-energy (Loc. 6) conditions prevailed. Crinoids and dasycladean algae identify stenohaline settings likely in shallow water of <30 m (Mamet et al. 1984; Berger and Kaever 1992; Flügel 2010), and local Girvanella-bearing oncoids identify a gently agitated environment with low sedimentation rates (Flügel 2010). Rhythmic interstratification of shale and carbonate illustrates episodic smothering of carbonate production. Potential controls are discussed below.

The biotite- and zircon-bearing whitish clay forming part of the upper recessive interval capping the L’Orignal Formation at Loc. 1 is interpreted to be altered pyroclastic sediment. Its mineralogy, including abundant euhedral phenocrysts of biotite and zircon, and a derived calc-alkaline magmatic source are attributes found in numerous altered
ancient ash falls preserved in the Turinian–Chatfieldian succession of eastern North America (Huff, 2008; Sell et al., 2015).

Geologic Age

At Loc. 3, the L’Orignal Formation contains “transitional” Blackriveran–Rocklandian macro- and microfaunal assemblage (Fig. 2.3; Wilson 1921; Barnes 1967) as well as the first appearance of the brachiopod *Triplesia cuspidata*, an indicator of the early Rocklandian (Kay 1937, 1942; Cameron and Mangion 1977). At Loc. 2, ~33 km to the east, an entirely Blackriveran conodont assemblage occurs in the equivalent succession (Barnes 1964, 1967). This appears to define a marked diachronocity of the formation base, younging to the west.

From an abundant aliquot of zircon crystals (Fig. 2.8e) recovered from the clay bed at Loc. 1, six elongate to equant multi-facetted (euhedral) chemically abraded crystals were selected to determine a U–Pb ID–TIMS date. Due to the high abundance of $^{206}\text{Pb}$ relative to $^{207}\text{Pb}$ in Phanerozoic zircon, the $^{238}\text{U}$–$^{206}\text{Pb}$ isotopic system is commonly considered the most robust age result (cf., Kamo et al. 1995). This produced a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ date of 453.36 ± 0.38 Ma (2σ, N = 6, and a mean square of weighted deviates (MSWD) = 0.52) (Fig. 2.9d) for the uppermost L’Orignal Formation.

Rockland Formation (revised)

Lithofacies succession

Whether as conceived by Raymond (1914) or as revised here, the lithostratigraphic characteristics of the Rockland Formation are geographically isolated (Fig. 2.7d). At its
type section (Loc. 3), the Rockland Formation (revised) retains the characteristic rhythmic alternation of thick beds of mostly lime mudstone (Raymond, 1914), with rare skeletal packstone and grainstone, and thin (3-5 cm), not thick-bedded calcareous shale beds (Figs. 2.5 and 2.10a). Previous references to thick shale beds (Raymond, 1914; Wilson, 1921) are in fact recessive intervals of increasing (upward) argillaceous limestone. As revised, the formation basal contact is stylolitic. Laminated coarse-grained grainstone of the Hull Formation disconformably overlies the Rockland Formation at Loc. 3. The disconformity truncates the Guttenberg Isotope Carbon Excursion, or GICE (Fig. 2.3), and Hull strata contain the Belodina confluens biozone (Barta et al. 2007).

**Depositional Environment**

The facies and macrofossil assemblage (Wilson, 1921) of the Rockland Formation represents a normal-marine low-energy (muddy) shallow-subtidal setting. Spatially discrete bodies of this lithofacies succession (Fig. 2.7d) occur coeval with deeper water sediment (see below), and suggest that this facies succession represents low-energy deposition along the top of a carbonate bank. Rare gravel-sized skeletal material, therefore, may identify episodic storm activity or shallowing that established elevated wave and (or) current regimes and rhythmic alternation with shale beds attest to periods of increasing siliciclastic influx resulting in temporary smothering of carbonate production and accumulation.
Fig. 2.10: Field characteristics of the Rockland and Hull formations. (a) Typical stratigraphic fabric of the Rockland Formation: thick beds of lime mudstone grading through argillaceous lime mudstone (beneath hammer) to recessive thin shale interbeds. Type section, Loc. 3. (b) Large-scale cross-stratification (highlighted in black) of Unit H1 (Hull Formation) at Loc. 1, confined to within 8 and 13 m above the base of the formation. (c) Typical alternating fossil (crinoid, bryozoan)-rich rudstone (a) with interbeds of bryozoan-rich and burrowed shale (b) for Unit H1 (Hull Formation). Loc. 1, scale bar in centimetres. (d) Irregular bodies of fine-grained grainstone with shale partings forming Unit H2 (Hull Formation) at Loc. 4. Metre stick for scale (e) Thick, amalgamated beds of coarse-grained crinoidal limestone forming Unit H3 (Hull Formation). Loc. 1, pencil for scale. (f) A shale bed separates Units H1 (A) and H3 (C) of the Hull Formation at Loc. 1, and contains an interpreted bentonite layer (arrow) displaying synsedimentary deformation. Pocket knife for scale.
Hull Formation (revised)

Raymond (1913) referred to the Hull beds from a large active quarry (Loc. 18), now flooded. As part of the Hull Formation (Uyeno, 1974), these beds were separated into a lower member of thin- to medium-bedded and fine- to medium-grained skeletal-rich (crinoidal) limestone with shale partings, and an upper member of thick-bedded (<3 m) medium- to coarse-grained skeletal-rich limestone, often cross-stratified. Conodont assemblages identified a late Blackriveran through Kirkfieldian affinity, with strongest correlation being with the Kings Falls Limestone in New York State (Uyeno 1974). In the Ottawa Embayment, the Hull Formation is conformably overlain by the Verulam Formation (Uyeno 1974). Exposures at Locs. 1, 6, and 9 were not available at the time of this earlier work, and reveal a lithic succession below Uyeno’s (1974) lower Hull Formation and above our proposed L’Orignal Formation. We refer it to the Hull Formation given similar stratigraphic motif and composition, thereby creating a tripartite (Units H1–H3) stratigraphy (Figs. 2.4 and 2.5; Table 2.1) wherein Units H2 and H3 are equivalent to Uyeno’s (1974) lower and upper members, respectively.

Lithofacies Succession

Unit H1 disconformably overlies the L’Orignal Formation (Fig. 2.8b), and is interpreted to be coeval with the revised Rockland Formation (Fig. 2.4). Thickness exposed at 5 localities (i.e., Locs. 1, 4, 5, 6, and 9) ranges from ~1.4 m to ~50 m and underlies a narrow (~25 km) geographic corridor extending along the embayment (Fig. 2.7d). At Locs. 1, 5, and 6, the succession consists of thin- to medium-bedded skeletal-rich rudstone and grainstone (with abundant crinoids and treptostomid bryozoans) interbedded
with fossiliferous burrowed dark calcareous shale with locally abundant cryptostomid bryozoans (Figs. 2.10b and 2.10c; Table 2.1). Limestone beds are massive to normally graded, with local crinoid holdfasts encrusting bedding tops. Quarry-wall exposures reveal a gentle (∼10°) southward depositional dip of discontinuous, locally overlapping beds. Large-scale cross-stratification with a southward plunge of similar magnitude occurs between 8 and 13 m above the formation base (Fig. 2.10b). At Loc. 4, the equivalent succession contains a rhythmic stratigraphy of an intermediate body of thicker-bedded/gravel-sized grained limestone-shale succession bounded by thinner bedded/ sand-sized limestone-shale successions (Fig. 2.4). At Loc. 20 (Blackburn quarry), the equivalent succession, packstone-shale succession (Salad Hersi, 1997), is greatly expanded, occupying tens of metres of thickness. In the western embayment (Loc. 9), in contrast, Unit H1 is only 1.4 m thick and contains, overall, a sand-sized mud-rich carbonate fraction.

Unit H2 is visually prominent due to a polygonal fabric (20-30 cm diameter) of fine-grained skeletal (echinoderm) grainstone bounded by fossiliferous and burrowed dark shale laminae to partings (Fig. 2.10d). Flame structures (Collinson 1994) demonstrate differential mechanical compaction of the limestone and synsedimentary deformation of intervening shale. The unit is recognized at Locs. 4, 5, and 9, and it ranges in thickness up to 26 m (Figs. 2.4 and 2.7).

Unit H3 consists mostly of medium to very thick (metre-scale) beds of coarse-grained echinoderm (crinoid) grainstone to rudstone (Fig. 2.10e), with amalgamated and cross-stratified bedding that exhibit abrupt to erosional tops (Uyeno 1974; Kiernan 1999). The unit occurs across the embayment (Fig. 2.7; Kiernan 1999), but maximum thickness is
uncertain due to post-Ordovician erosion. The unit abruptly overlies either Unit H1 or H2. At Loc. 1, a thin (5 cm) dark-grey shale separates Units H1 and H3, and contains a thin (<2 cm) interbed of sticky whitish clay (Fig. 2.10f) similar to that found capping the L’Orignal Formation at this same site. Less common facies associated with Unit H3 are lime mudstone, shale, and skeletal packstone. At Locs. 5, 9, and 20, enrolled and contorted grainstone-shale stratigraphy forms intervals of <1 m thickness, bounded by undisturbed planar bedding.

*Depositional Environments*

Sedimentological features of Unit H1 identify episodic accumulation of coarse-grained carbonates along a low-gradient slope, and subject to fluctuating transport energy and influx of siliciclastic fines. The fossil assemblage is similar to the open-platform biofacies of the lower Trenton Group in the northern Appalachian Basin (Titus and Cameron, 1976). Paleogeographically, Unit H1 lies adjacent to the interpreted low-energy bank facies of the Rockland Formation (revised), creating a regional depositional mosaic of banks and large interbank channels or micro-seaways similar to muddy inner shelf-margin wedges and skeletal-rich channels inboard of the Florida Keys (Enos, 1977; Wanless et al., 1995). Rare coarse-grained limestone in the Rockland Formation could be related to storm-induced transport from the adjacent channel (cf. Ball et al., 1967; Tedesco and Wanless, 1995).

Local upsection transition from Unit H2 to H3 suggests the two facies are genetically related: the former representing a quieter, deeper water periphery to high-energy shallow-water shoals (Uyeno 1974). Absence of a similar transition between Units H1 and H3,
and the disconformable superposition of the latter on Rockland strata at Loc. 3, are interpreted as evidence for lateral expansion of high-energy shoal systems across paleosurfaces capping older depositional systems. Lower energy facies that locally accompany Unit H3 in the western embayment may illustrate quieter intershoal areas or define a higher order stratigraphy of base-level change. Synsedimentary deformation features in Unit H3 are similar to paleoseismites in equivalent Chatfieldian strata in the eastern United States (Pope et al., 1997; Jewell and Ettensohn, 2004).

**Lowville to Napanee lithostratigraphy, Southern Ontario**

We present a composite section of the Lowville through lower Napanee succession from the northern Appalachian Basin (southern Ontario) (Fig. 2.1), and correlation with what appears to be the remains of the Selby type section (Fig. 2.11), to better facilitate lithostratigraphic correlation with the Ottawa Embayment. Our interpretation differs, in part, from previous work that addressed parts of the composite section (Cameron, 1968; Cameron and Mangion, 1977; Conkin, 1991; McFarlane, 1992; Cornell, 2001).

We recognize a similar lower and upper division of the Lowville Formation, but with the added appearance of stromatolites (McFarlane, 1992) that form biostromes in the lower division and small to large (metre-scale) mounds (Fig. 2.12a) in the upper division beds that contain a profusion of fragmented Tetradium. Our interpreted Lowville-Watertown boundary, positioned lower than that of McFarlane (1992), displays minor to dramatic depositional relief (< tens of cm) associated with burial of upper-division stromatolite mounds beneath thickly bedded Watertown strata (Fig. 2.12a). The latter formation exhibits, overall, thick beds of muddy limestone intercalated with at least 3 intervals of
Fig. 2.11: A composite lithostratigraphic section of the Pamelia through lower Napanee formations in the Kingston–Napanee region of the northern Appalachian Basin (see Fig. 2.1 for locations of sections), and correlation with the type section of the Selby Formation (Loc. 14) in this same region. Illustrated are the interpreted equivalent lower and upper divisions of the Lowville Formation, and prominent basal grainstone beds (green) associated with the Selby and Napanee formations. The photograph illustrates a lens of whitish crystalline claystone host to abundant silt-size angular skeletal fragments (a) overlain by well sorted crinoidal-skeletal grainstone (b) with clasts differentially Fe-oxized. This is an interpreted reworked volcanic ash, its stratigraphic position being very similar to the Millbrig bentonite in the Ottawa Embayment.
**Fig. 2.12:** Field characteristics of the upper Blackriveran – lower Trentonian succession in the Kingston–Napanee region. (a) Boundary (white arrow) separating the equivalent lower (L1) and upper (L2) divisions of the Lowville Formation at Loc. 16a. A discontinuous clay bed, likely a bentonite, occurs along the contact. Illustrated in this exposure are three stromatolite mound complexes (s) in the upper division of otherwise thinly bedded Tetradium-bearing limestone. The mounds are differentially overlain (black arrow) by a thick bed of strongly burrowed Watertown Formation (w). (b) Equivalent lithofacies W2 (a) of the Watertown Formation at Loc. 13 is interbedded with two intervals of shaley fossiliferous lime mudstone (arrows) otherwise similar to Kay’s (1929) Glenburnie Shale. (c) Disconformable contact at Loc. 13 between lithofacies W2 of the Watertown Formation (w) and the Selby Formation (s). Hammer for scale. The arrow denotes a very thin (<2 cm) skeletal grainstone that is interpreted as a marine transgressive deposit forming the base of the Selby Formation (see Fig. 2.11). (d) Selby strata (S) at the remains of the formation’s type section (Loc. 14), overlain by thinly bedded coarse-grained crinoid grainstone (t) that is an interpreted marine transgressive deposit forming the basal Napanee Formation (Fig. 2.11). Metre stick intervals are 10 cm. (e) Lower Napanee Formation at Loc. 13, in ascending order: basal transgressive deposit (t); 10-cm-thick clay bed (b) with reworked clasts of quartz arenite, siltstone, fossiliferous crystalline claystone (similar to texture illustrated in Fig. 2.11), and abundant fossil fragments; and succession (double arrow) of interbedded shale and packstone to grainstone. Hammer (right) for scale.
Kay’s (1929) “Glenburnie Shale” (Fig. 2.11 and 2.12b; see also Cameron, 1968; Cameron and Mangion, 1977; Conkin, 1991). The upper part of the formation exhibits a similar W2-like facies (Fig. 2.12b) including locally bored hardgrounds, one of which defines the formation top (Fig. 2.12c; McFarlane, 1992). Above this paleosurface, the basal deposit of the Selby Formation is a thin (<2 cm) crinoidal grainstone (Figs. 2.11 and 2.12c), succeeded by the hackly weathering, fossiliferous, and argillaceous lime mudstone characteristic of this unit (Kay, 1937). A difference of ~2.5 m in formation thickness occurs between Locs. 13 and 14, a distance of ~14 km, constrained by a basal crinoidal grainstone of the Napanee Formation (Figs. 2.11 and 2.12d). At Loc. 13, the lower part of the grainstone contains lenses to partings of whitish skeletal microbioclastic wacke consisting of fine-grained (~30-140 µ) skeletal fragments disseminated in a matrix of white crystalline claystone (Fig. 2.11). The remaining grainstone contains three prominent attributes: strongly abraded skeletal allochems that display varying amounts of Fe-oxide alteration; intraskeletal (bryozoan, gastropod) paleoporosity filled with micrite; and, rare, rounded, sand to gravel-size lithoclasts of skeletal- and lithoclast-bearing whitish and Fe-oxidized crystalline claystone. Above the grainstone is a 10-cm-thick clay bed (Fig. 2.12e) host to similar claystone lithoclasts along with siliciclastic (arenite, siltstone) lithoclasts and individual fossil fragments of which specific fossil types are similar to those of the lower Trenton Group of New York state (Cameron 1968; Titus and Cameron 1976). The remainder of the lower Napanee Formation consists of interbedded shale and thinly bedded limestone (packstone, grainstone), with local amalgamated bedding (Fig. 2.12e).
Discussion

Upper Turinian – lower Chatfieldian lithostratigraphy of the Ottawa Embayment records three stages of carbonate-platform development (Fig. 2.13). When compared to equivalent lithostratigraphy in the northern Appalachian and western Quebec basins, the intermediate Stage 2 characterizes embayment-specific stratigraphic patterns. The bentonite within the uppermost L'Orignal Formation of the Ottawa Embayment has a similar stratigraphic position as the Millbrig bentonite in northern New York State (Sell et al., 2015). Our age (453.36 ± 0.38 Ma) date is within the 2σ uncertainty of the best-age determination (452.86 ± 0.29) reported for this ancient ash-fall deposition in the United States (Sell et al., 2013). Our older mean age, however, may establish a more robust geochronological framework (Fig. 2.13) for upper Turinian – lower Chatfieldian stratigraphy (Fig. 2.13) when integrated with the age of Katian–Sandbian boundary (Cooper et al., 2012) and the Deicke bentonite (453.74 ± 0.20; Sell et al., 2013).

Platform Stages and Inter-regional Correlation

Stage 1

Traced regionally among the three basins, the Lowville Formation characterizes a once expansive shallow-water carbonate platform of varied high- and low-energy environments (Harland and Pickerill, 1982; Salad Hersi and Dix, 1999; Brett et al., 2004). The lower-upper division traced from the Ottawa Embayment into the northern Appalachian Basin (southern Ontario) is accompanied by appearance of stromatolites (Figs. 2.11 and 2.12a). Moving farther south, the Weaver Road Beds (WR, Fig. 2.13) in New York state (Brett et al., 2004) appear to occur in a similar stratigraphic position as
**Fig. 2.13:** Upper Turinian – lower Chatfieldian stages of platform development and nature of base-level change associated with paleosurfaces (1–5) in the Ottawa Embayment. This is compared with depositional sequence frameworks for the eastern United States (sequence 1, Sell et al. 2015 and sequence 2, Brett et al. 2004) and related base-level change associated with a composite Mohawkian section of the northern Appalachian Basin, and interpreted equivalent changes in the western Quebec Basin (based on Harland and Pickerill 1982). Compilation of regional bentonite stratigraphy in the central to eastern United States is from Sell et al. (2015), as is the vertical distribution of biozones relative to the Mohawkian section and stage boundaries. Three geochronological reference horizons include the base of the Katian stage (Cooper et al., 2012), the Millbrig (M) bentonite (this study), and the Deicke (D) bentonite (Sell et al., 2013) forming a rescaled chrono/biostratigraphic framework between these data points. Abbreviated older stage names include: Ro, Rocklandian; Kirk, Kirkfieldian; and Sher, Shermanian; and are fit to the Mohawkian lithostratigraphy based on Sweet and Bergström (1971). Lower lithostratigraphy of the Ottawa Embayment is fit to the Mohawkian standard section on the basis of bentonite stratigraphy, whereas there is less certainty associated with the Hull Formation, the upper part tied to conodonts biozones (Uyeno, 1974; Barta et al., 2007). Abbreviated conodont zones refer to the North Atlantic (left) and Midcontinent (right) zones: *Amorphognathus tvaerensis, A. superbus, Phragmodus undatus, Plectodina tenuis, Belodina confluens*, and the abbreviated graptolite biozones refer to the Taconic foreland (left) and international graptolite zones (right): *Climacograptus bicornis, Corynoides americanus; Or, Orthograptus ruedemanni, Diplacanthograptus caudatus.*
the upper division of the Lowville Formation. If this correlation is correct, it defines a regional (200+ km) environmental gradient from Tetradium-rich facies in the embayment (and extending east into the western Quebec Basin; Okulitch, 1935; Harland and Pickerill, 1982) to only microbial production in the northern Appalachian Basin (New York). In paleogeographic coordinates (Torsvik and Cocks, 2017), increased microbial production extends into a more offshore setting. This likely reflects regional changes in salinity, nutrient loading, and (or) temperature related to platform circulation and (or) slight variation in bathymetry. However, an apparent rise in water depth across the lower-upper division of the formation in the embayment versus shallowing characterized by appearance of the Weaver Road microbial mounds (Brett et al., 2004) might also reflect differential subsidence creating more open circulation toward the embayment and Quebec Basin.

The end of Stage 1 in the Ottawa Embayment is characterized by a post-Lowville disconformity (paleosurface 1, Fig. 2.13), certainly locally developed (Fig. 2.6c). In the northern Appalachian Basin (southern Ontario), our interpreted Lowville–Watertown boundary is a flooding surface with microbial mounds buried beneath burrowed Watertown lime mudstone (Fig. 2.12a). Farther south, the equivalent paleosurface is interpreted as a sequence boundary following forced regression (Brett et al., 2004). Variation in the nature of this regional paleosurface suggests the influence of local, not regional base-level controls.
Stage 2

This stage defines a succession (stages 2a-2c) of stratigraphic attributes in the embayment distinct from those in adjacent basins:

**Stage 2a**

Thickly bedded muddy facies of the Watertown Formation and equivalent Leray Formation in the western Quebec Basin can be traced regionally. In the Ottawa Embayment, however, this facies postdates an initial ravinement during west-directed translation of the shoreline and deposition (facies W1; Table 2.1) across the post-Lowville paleoplatform (paleosurface 1, Fig. 2.13), and it is accompanied by net stratigraphic condensation in the western embayment compared with thicker deposits in the east (Fig. 2.6b). This latter contrast is interpreted to document elevated subsidence to the east. Even thicker (35 m), but local, equivalent successions occur in the western Quebec Basin, although these have been regarded as possibly the result of error in placement of formation boundaries (see Harland and Pickerill, 1982). As the thickness of the Watertown Formation in the western embayment is similar to that of the northern Appalachian Basin, eastward thickening in the embayment defines a preferential axis of differential subsidence extending from the Ottawa Embayment into the western Quebec Basin.

In the Ottawa Embayment, a post-Watertown disconformity (paleosurface 2, Fig. 2.13) marks erosion associated with a drop in base level. The paleosurface was mantled, first, by an interpreted volcanic-ash deposit that may be the Deicke deposit on the basis of similar stratigraphic position as found in the northern Appalachian Basin (Fig. 2.13). The
ancient ash deposit in the embayment, however, was reworked during accumulation of shoreface siliciclastics. Compared with bounding carbonate strata, the siliciclastics document a drop in base level. In contrast, the equivalent paleosurface in the northern Appalachian Basin is a bored marine hardground in southern Ontario (Fig. 2.12c) and, farther south, part of a net rise in base level in northern New York (Brett et al., 2004). In the western Quebec Basin, an apparent conformity with the overlying Trenton Group strata (Harland and Pickerill, 1982) further demonstrates the local nature of base-level change in the Ottawa Embayment.

Stage 2b

Marine transgression yielding the L’Orignal and equivalent Selby successions in the Ottawa Embayment and northern Appalachian Basin (southern Ontario), respectively, was associated with local development, at least, of coarse-grained transgressive deposits (Fig. 2.13). In the Quebec Basin, equivalent transgression may have resulted in formation of a stratigraphically condensed interval defined by the lower Trenton Group, as indicated by several co-related features: a thin (3-6 m) succession of alternating thinly bedded bioclastic micritic limestone and shale that identifies episodic smothering of carbonate accumulation; episodic high-energy modification of the seafloor; extensive, and top-down, bioturbation denoting periods of little or no sediment accumulation (Harland and Pickerill, 1982; Lavoie, 1995; Mehrtens, 1988); and a mixed late Turinian – early Chatfieldian macrofossil assemblage (see Lavoie, 1995).

The rhythmic carbonate–siliciclastic stratigraphic fabric of the L’Orignal Formation identifies greater affinity to this latter stratigraphic motif than the more homogenous
admixing of siliciclastics in the Selby Formation in the northern Appalachian Basin (southern Ontario). Given the appearance of shoreface siliciclastics in the lower L’Orignal Formation, a rhythmically intercalated siliciclastic carbonate stratigraphy might represent distal deposits of a shore-based siliciclastic source timed with base-level drop or increased riverine input due to paleoclimatic variation. Alternatively, they reflect episodic incursion of marine shale during periods of sea-level highstand demonstrating greater sedimentary continuity with the western Quebec Basin.

A widespread but variably developed post-Selby erosional vacuity (paleosurface 3, Fig. 2.13) is recognized across the northern Appalachian Basin (Cameron, 1968; Cameron and Mangion, 1977; Mitchell et al., 2004) and illustrated in the southern Ontario region by differential (~2.5 m) erosion between Locs. 13 and 14 (Fig. 2.11). This has been correlated regionally with the M5A sequence boundary by Sell et al. (2015), whereas Brett et al. (2004) interpreted the formation top in northern New York state as a maximum flooding surface. In the Kingston–Napanee region, post-Selby erosion predates a marine transgressive grainstone, and it remains uncertain if the differential erosion (Fig. 2.11) is related to either a base-level drop or marine ravinement, or both. Quartz arenite lithoclasts in the claystone bed above this basal grainstone at Loc. 13 attest to a siliciclastic source not defined in outcrop and, therefore, may serve as evidence for base-level drop. Lenses of skeletal-rich crystalline claystone in the lowermost part of the grainstone (Fig. 2.11), and presence of altered claystone lithoclasts, also suggest the reworking of an altered ash deposit that may have been within the upper Selby, and equivalent to the Millbrig bentonite. In the Ottawa Embayment, the equivalent disconformable surface shows no evidence for base-level drop but, instead, a rise.
Stage 2c

Following accumulation of the Millbrig bentonite in the Ottawa Embayment, lateral compartmentalization of lithofacies produced a paleodepositional mosaic of shallow carbonate banks and deeper, higher energy, channels or micro-seaways (Fig. 2.7d). In the northern Appalachian Basin, alternating skeletal-rich carbonate and shale of the Napanee Formation (Fig. 2.11; Cameron and Mangion, 1977) characterize a regional deepening above an interpreted marine hardground paleosurface capping the Selby Formation (Brett et al., 2004). In the western Quebec Basin, the equivalent paleosurface would lie within the interpreted Rocklandian condensed succession (Fig. 2.13). By late Rocklandian, high-energy cool-water carbonate deposition began to appear across paleosouthern Laurentia (Brookfield, 1988; Lavoie, 1995; Holland and Patzkowsky, 1996; Pope and Read, 1997, 1998). In the Ottawa Embayment, formation of the carbonate bank-seaway mosaic may have restricted cooler waters and biota migration to the deeper water corridors.

Stage 3

In the Ottawa Embayment, this stage is defined by shallowing (across paleosurface 4; Fig. 2.13) and expansion of high-energy crinoidal shoals and intershoal deposits associated with Units H2 and H3 of the Hull Formation. Uyeno (1974) considered the Hull Formation not to be younger than the middle Kirkfieldian (Fig. 2.13), and part of an inter-regional distribution of high-energy carbonate ramp systems encircling the regional foreland interior as represented by the Deschambault Formation (Quebec Basin; Lavoie, 1995) and Kings Falls Limestone (northern Appalachian Basin; Brett and Baird, 2002) (Fig. 2.13). In the northern Appalachian Basin, onset of these coarse-grained deposits is
correlated with a base-level fall (M5B sequence boundary, Fig. 2.13; Brett et al., 2004; Sell et al., 2015). In the Ottawa Embayment, a prominent hiatus separates the Rockland and Unit H3 strata of the Hull Formation at Loc. 3 as defined by the presence of the Belodina confluens conodont biozone in the latter strata and erosion of the upper part of the GICE in the former (Fig. 2.3). Deposition of Unit H3 on Unit H1 strata at Loc. 1 also demonstrates a base-level fall defined by abrupt shallowing of depositional conditions.

**Controls on Platform Development, Ottawa Embayment**

Third-order depositional sequences comprise the Upper Ordovician foreland succession through the eastern United States (Holland and Patzkowsky, 1996). With ongoing improved stratigraphic resolution, this approach has provided two interpreted depositional sequence successions of the upper Turinian – lower Chatfieldian interval within the northern Appalachian Basin (Fig. 2.13; Brett et al., 2004; Sell et al., 2015). Details of sequence stratigraphic framework for the Ottawa Embayment remain an ongoing study (see Chapter 4). Here, we propose that Stages 1 and 3 of platform development in the Ottawa Embayment (Fig. 2.13) have general facies attributes in common with regionally expansive depositional systems along the Taconic foreland interior. The net succession records regional paleoceanographic transformation from warm- to evolving cool-water carbonate deposition (e.g., Brookfield, 1988; Keith, 1988; Lavoie, 1995; Pope and Read, 1997; Holland and Patzkowsky 1998) timed with or following increased regional subsidence (Ettensohn, 2008) allied with apparent north-directed subduction beneath Laurentia (Macdonald et al., 2017).
Stage 2, in contrast, identifies temporary emergence of stratigraphic patterns specific to the Ottawa Embayment (when compared to adjacent basins), and is interpreted to represent local structural influence. This is reflected by several independent features that identify intrabasinal differential subsidence: (i) the along-axis differences in thickness and stratigraphic condensation of the Watertown Formation, (ii) the local post-Watertown base-level drop compared to apparent rise in adjacent basins (Fig. 2.13), and (iii) platform segmentation into banks and deeper water micro-seaways following the Millbrig ash deposit. Differential subsidence is also expressed by anomalous thickening of formations proximal to the traces of present-day regional Gloucester and Rideau faults (Fig. 2.7).

These are long-lived structures (Rimando and Benn, 2005), with evidence of structurally influenced local erosion with onset of foreland tectonism in the Middle Ordovician (Dix and Al Rodhan, 2006), anomalous thickening of the Blackriveran Pamelia Formation (Salad Hersi and Dix, 1999), and local micrograben development contemporary with Late Ordovician (Edenian) collapse of the carbonate platform (Dix and Jolicoeur, 2011).

Anomalous thickening of the Lowville Formation only proximal to the Gloucester and Rideau faults (Fig. 2.6a) serves to demonstrate local structural influence within a regional framework of deposition (see above).

The origin of local synsedimentary structural activity along this part of the Taconic foreland interior is interpreted in context of the far-field response of inherited structurally weakened continental crust (in this case, the axis of an aborted late Precambrian (590–570 Ma) rift: Burke and Dewey 1973; Kumarapeli, 1985; McCausland et al., 2007) to changing dynamics of distal plate-boundary tectonism (Marshak and Paulsen, 1998). Paleostress-field analyses have shown that movement along regional faults in the
embayment (Fig. 2.1), as well as evidence for intracrystalline deformation in Ordovician carbonate rocks in the northern Appalachian Basin, are compatible as far-field responses to periods of distal Paleozoic orogenesis along the Laurentian margin (Craddock and van der Pluijm, 1989; Rimando and Benn, 2005). Localization of structural activity in the foreland interior is enhanced by an along-axis variation in rheological properties of the crust (Fernández and Ranalli, 1997). We note that repeated local thickening of formations near the intersection of the Gloucester and Rideau faults may be also controlled by differential subsidence of a geophysically defined isolated block of the Morin-Adirondack Highlands basement terrain (Fig. 2.1). Local faulting of Chatfieldian age is documented elsewhere along the foreland interior (Bradley and Kidd, 1991; Lavoie, 1995; Jacobi and Mitchell, 2002) in response to foreland flexure (Bradley and Kidd, 1991).

There may also be a record of far-field structural control on base level in the embayment timed with two periods of increased regional density of bentonite occurrences (Fig. 2.13). From the dataset of Sell et al. (2015), which documents bentonite stratigraphy over ~2.5 × 106 km² in the central to eastern United States, the first period brackets onset of interpreted differential subsidence (Stage 2a) in the Ottawa Embayment (Fig. 2.13). The second period defines renewed elevated frequency of volcanic ash fall, marked by the regional Millbrig bentonite, timed with platform segmentation (Stage 2c) in the embayment. Along plate boundaries, large magnitude earthquakes can trigger co-related eruptions (Linde and Sacks, 1998; Nishimura, 2017) and trigger remote earthquakes within plates and along extensions of active and aborted continental rift axes (Hill et al., 1993; Mohamad et al., 2000; Hough et al., 2003). The source of the Millbrig bentonite
was likely proximal to the Laurentian margin (Macdonald et al., 2017), the erupted mass among the largest known in the Phanerozoic (Huff et al., 1996; Mason et al., 2004). For the Ottawa Embayment, timing of this ancient ash fall prior to the interpreted platform segmentation yielding banks and micro-seaways might be viewed as mere coincidence given an elevated frequency of ash fall (cf. Miall, 1992). However, the record of platform segmentation coincides with an interpreted abrupt regional deepening in the northern Appalachian Basin (Brett et al., 2004) within this period of distal change in plate-boundary dynamics (Macdonald et al., 2017). In other basins, more dramatic examples of platform segmentation yielding banks and intraplatform seaways spatially allied with inherited structure have been interpreted incorporating similar far-field responses (e.g., Eberli and Ginsburg, 1987; Masaferro and Eberli, 1999; Bachetal et al., 2004). Based on this comparative work, subsequent expansion of the Hull Formation as part of an inter-regional sedimentary cover can be interpreted as rates of sediment production and (or) supply matched then exceeded rates of differential subsidence.

Conclusions

A revised lithostratigraphic framework is presented for the upper Turinian – lower Chatfieldian (Blackriveran–Kirkfieldian) stratigraphy in the Ottawa Embayment. The proposed L’Orignal Formation (early Rocklandian in age) is equivalent to the lowermost Rockland Formation (now revised), and lithostratigraphically equivalent to, but lithologically distinct from, the Selby Formation in the northern Appalachian Basin. The Hull Formation is revised, its lower part considered a facies equivalent to the revised Rockland Formation.
The lithostratigraphic succession across the embayment defines three stages of carbonate-platform development. Compared with contemporary stratigraphy in adjacent basins, facies patterns of inter-regional extent (Stages 1 and 3) bound the intermediate Stage 2 that preserves embayment-specific stratigraphic patterns. Stratigraphic attributes of Stage 2 include, in succession, dramatic east-directed differential thickening and west-directed stratigraphic condensation, post-Watertown base level fall (in contrast to base-level rise in adjacent basins) defined by appearance of shoreface siliciclastics, and lateral facies portioning across the embayment producing a paleodepositional mosaic of low-energy carbonate banks bounding intraplatform micro-seaways. The latter event postdates accumulation of the regional Millbrig bentonite, preserved locally in the embayment and dated at 453.36 ± 0.38 Ma. Stages 1 and 3 form local expressions of a net regional paleoceanographic transformation across paleo-southern Laurentian: late Turinian warm-water carbonates to early Chatfieldian cooler carbonate production. Stage 2 defines a structural transition between these states within the embayment and was contemporary with changing subsidence rates along the Laurentian margin. This history documents a far-field response localized in this part of the foreland interior due to predisposed structural weakness along the paleoaxis of a late Precambrian fault system.
CHAPTER 3: δ^{13}C STRATIGRAPHY OF A TURINIAN-CHATFIELDIAN (UPPER ORDOVICIAN) FORELAND SUCCESSION, OTTAWA EMBAYMENT (CENTRAL CANADA): RESOLVING LOCAL AND INTER-REGIONAL ISOTOPE EXCURSIONS IN A TECTONICALLY ACTIVE BASIN

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Abstract

Positive $\delta^{13}$C$_{\text{carb}}$ excursions are correlated through an upper Turinian to lower Chatfieldian carbonate-platform succession along the axis of the Ottawa Embayment and into outliers of the northern Ottawa-Bonnechere graben in central Canada. Successive Turinian excursions (E1 and E2) are lithostratigraphically constrained by erosional surfaces and hosted within the Watertown and overlying L’Orignal formations, respectively, the latter coeval with the Selby Formation in the adjacent northern Appalachian Basin. The excursions coincide with periods of regional transgression, but geographic patterns of $^{13}$C depletion versus enrichment coincide with structurally defined areas of stratigraphically condensed and preferentially thickened formation successions, respectively. Differential subsidence is interpreted to have created bathymetric variation resulting in intrabasinal restriction of seawater exchange between these areas, with preferential $C_{\text{org}}$ recycling with stratigraphic condensation. By early Chatfieldian time, segmentation of the once regional carbonate platform (L’Orignal Formation) produced a regional mosaic of low-energy muddy carbonate banks (Rockland Formation) and a deeper water platform (lower Hull Formation) settings subject to fluctuating high to low energy current flow. Excursion E3 occurs in both successions, but $^{13}$C enrichment is associated only with the bank-top muddy facies. This may identify preferential photosynthetic drawdown of $^{12}$C across the bank tops due to limited seawater exchange across the bank-deeper platform boundaries. Excursions E1 to E3, and a younger excursion (E4) in the Hull Formation, are correlated with varying confidence with excursions across southern Laurentia, excursion E3 being the local expression of the Guttenberg $\delta^{13}$C excursion. Our study supports local modulation of regional, if not
global, $\delta^{13}C$ excursions arising from structurally controlled changes in oceanography and productivity.

**Introduction**

Positive $\delta^{13}C_{\text{carb}}$ isotope excursions are documented from Turinian-Chatfieldian (Upper Ordovician) strata in North America (Hatch et al., 1987; Ludvigson et al., 2004; Saltzman and Young, 2005; Young et al., 2005; Barta et al., 2007; Bergström et al., 2010b; Metzger et al., 2014). Of these, the early Chatfieldian Guttenberg $\delta^{13}C$ excursion (Hatch et al., 1987) or GICE is also recognized from sites in Scandinavia, Estonia, Malaysia, and China (Kaljo et al., 2004; Bergström et al., 2010a, b). Of short duration (<500,000 yrs; Ludvigson et al., 1996), this event forms an important inter-regional, maybe global, correlation tool (Bergström et al., 2009; Cooper et al., 2012). Noted geographic variation in $\delta^{13}C_{\max}$ values for the GICE across the United States have been interpreted from two different perspectives: (1) a variable signal due to regional to local oceanographic and related environmental gradients (Young et al., 2005; Panchuk et al., 2006); and (2) a common regional signal produced from an isotopically homogenous marine reservoir with variation caused by local syndepositional, but especially post-depositional diagenesis (Metzger and Fike, 2013; Metzger et al., 2014). For the most part, previous work in North America documents excursions from relatively quiescent tectonic settings, parts of the regional Taconic foreland system and coeval intracratonic basins (Bergström et al., 2010b). We present a correlation of $\delta^{13}C$ profiles through upper Turinian-lower Chatfieldian strata in the Ottawa Embayment (central Canada): the foreland-interior succession is underlain by a Neoproterozoic intracratonic fault system, and subject to far-
field syndepositional tectonism in response to orogenic activity (Oruche et al., 2018, or Chapter 2). Geographic patterns in $\delta^{13}$C_carb (including the GICE), when integrated with lithostratigraphy and structure, reveal roles of local oceanographic and sedimentary modulation of widespread isotopic excursions. Our study suggests that both environmental gradients and stratigraphic-bound diagenesis are important considerations.

**The Ottawa Embayment**

**Geologic setting and structural history**

The Ottawa Embayment is underlain by lower Paleozoic (Cambrian through Upper Ordovician) strata, once contiguous with the epicontinental St. Lawrence Platform of southern Laurentia (Sanford, 1993). Its present-day indentation along the southern Canadian Shield (Fig. 3.1) reflects the effects of both erosion and faulting. The embayment is delimited from the western Quebec Basin to the east by the Oka-Beauharnois anticline, and from the northern Appalachian Basin (Ontario) to the southwest by the Frontenac Arch (Fig. 3.1; Sanford, 1993). Cambrian through Middle Ordovician (Darriwilian) strata form platform-interior facies of the cratonic Sauk Megasequence (Fritz et al. 2012; Lavoie et al. 2012). Middle through Upper Ordovician (Katian: Ka1) strata define a foreland-interior platform succession, the local expression of the cratonic Tippecanoe I Megasequence (Fritz et al. 2012). Subsequent platform collapse and expansion of a deep-water shale basin (Billings Formation) were driven by north-directed subsidence related to Taconic orogenesis (Lavoie, 1994). Subsequent shallowing during the later Katian (Ka2-Ka4) is recorded by net shallowing and deposition of an
Fig. 3.1: Geological elements of the Ottawa Embayment (OE) and outliers extending northwest along the axis of the Ottawa-Bonnechere graben (OBG). Section locations are numbered with geographic coordinates provided in Appendix A. The OBG fault distribution, with the Gloucester and Rideau faults highlighted, illustrates a west-directed curvature (from E-W to NW-SE). The OE is delimited from the adjacent western Quebec Basin (wQB) by the Oka-Beauharnois anticline, and from the northern Appalachian Basin (nAB) by the Frontenac arch (FtA). Precambrian geological components include the Grenville mafic dyke swarm (green), and the NW-SE-oriented distribution of the Proterozoic (Grenvillian) lithotectonic terrains: CMB, Central Metasedimentary Belt; FA, Frontenac–Adirondack lowland terrain; and MAH, Morin–Adirondack highland terrain. Possible extension of the MAH-FA boundary (dashed outline) beneath the OE stratal succession, and possible isolated block of the MAH terrain, is based on aeromagnetic geophysical patterns (Gupta 1991). Distribution of Lower Paleozoic megasequences (Dix 2012) subdivides the foreland Tippecanoe I Megasequence into an initial platform-interior, then basin-fill succession. Map is based on Baer et al. (1977), Sanford and Arnott (2010), and Bleeker et al. (2011), and modified from Oruche et al. (2018).
orogen-derived (Taconic) regional clastic wedge (Sanford, 1993). Faults within the embayment define the southeastern extension of the intracratonic Ottawa-Bonnechère graben (Fig. 3.1; Kay, 1942). The regional structure reflects ~600 million years of activity: initiation of a Neoproterozoic aborted rift; Early Paleozoic, Mesozoic, and Quaternary faulting; heating; and, post-Paleozoic uplift and erosion (Burke and Dewey, 1973; Crough, 1981; Rimando and Benn, 2005; Dix and Al Rodhan, 2006; McCausland et al., 2007; Ma and Eaton, 2007; Dix and Al Dulami, 2010; Dix and Jolicoeur, 2011; Roden-Tice et al., 2005, 2012; Dix et al., 2012; Hardie et al., 2017; Lowe et al., 2017).

**Turinian-Chatfieldian lithostratigraphy**

Figure 3.2 compares recently revised upper Turinian to mid-Chatfieldian lithostratigraphy in the Ottawa Embayment (Oruche et al., 2018) with older terminology as well as coeval successions in other sedimentary basins. In the embayment, Turinian formations (Pamelia through L’Orignal) record a net deepening-upward succession of warm, shallow-water, carbonate facies. Rhythmic intertidal-supratidal deposits of the Pamelia Formation give way to peritidal-shallow subtidal facies of the Lowville Formation. The latter formation culminates in development of *Tetradium* biostromes and mounds (Wilson, 1946; Salad Hersi and Dix, 1999; Oruche et al., 2018), part of an apparent regional environmental gradient passing into microbial buildups in the northern Appalachian Basin (Fig. 3.2; Oruche et al., 2018). The Pamelia and Lowville formations are anomalously thick (56 and 25 m, respectively) in the central embayment: the Pamelia Formation thins to ~36 m west of Ottawa, and is < 7 m along the periphery of the western Quebec Basin; the Lowville Formation maintains a relatively constant (7-8 m) thickness outside of the central region.
**Fig. 3.2:** Upper Turinian – lower Chatfieldian stratigraphic nomenclature for the study area (Ottawa Embayment) and northern Appalachian Basin (Ontario), shown in grey, and equivalent successions in adjacent basins referred to in the text. Previous interpretations of the position (a, b) of the Turinian-Chatfieldian (or older Blackriveran–Trentonian) boundary is illustrated, along with known or suspected distribution of selected bentonites. The hatched area related to the Selby Formation represents a prominent disconformity such that overlying reworked bentonite may be equivalent to the Millbrig bentonite in New York. The types of prominent biogenic buildups beneath and along the Lowville-Watertown formation boundary are indicated. Global chronostratigraphy is based on Ogg et al. (2008) and the use of the term megasequence follows Fritz et al. (2012). The positions of the bentonites are placed based on each author’s interpretation.
but thins to ~ 4-6 m in the western Quebec Basin (Harland and Pickerill, 1982; Salad Hersi, 1997; Salad Hersi and Dix, 1999). This underscores the likely role of differential subsidence related to the central embayment (Salad Hersi and Dix, 1999; Oruche et al., 2018). Possible structural controls include syndepositional activity along the intersection of the Rideau and Gloucester faults (Fig. 3.1), and differential movement of an apparent isolated Precambrian block of the Morin-Appalachian Highland terrain (a in Fig. 3.1; Oruche et al., 2018).

A disconformity capping the Lowville Formation is overlain by coarse-grained skeletal-rich carbonate interpreted as a basal transgressive succession of the Watertown Formation that thickens into the western embayment and appears locally along the northern limit of the embayment (Oruche et al., 2018). Upsection deepening established a lower energy, normal-marine subtidal facies, one associated with the thickly bedded (metre scale) limestone typical of the Watertown and equivalent successions in adjacent basins (Barnes 1967; Cameron, 1968; Cameron and Mangion, 1977; Oruche et al., 2018). Along the embayment axis, west-directed thinning of the formation (to < 4 m) coincides with increased stratigraphic condensation as indicated by an increase in abundance of hardgrounds and related omission surfaces.

The succeeding L’Orignal Formation disconformably overlies the Watertown Formation. The disconformity truncates burrows in the older formation that are filled with remains of a crinoid-bearing volcanic-ash carbonate-clay residue. Above this erosional surface is locally preserved shoreface phosphatic quartz arenite or, elsewhere in the region, quartz arenite lithoclasts occur within a basal skeletal-rich limestone facies and provide evidence that a basal arenite facies was once present (Oruche et al., 2018). The L’Orignal
Formation maintains a relatively constant thickness (~2 to 3 m) across the region, and displays a marked lateral variation in high- and low energy basal carbonate facies developed on top of the post-Watertown disconformity. The formation is interpreted to occupy the same lithostratigraphic position as the Selby Formation in the northern Appalachian Basin (Fig. 3.2; Cameron and Mangion, 1977; Oruche et al., 2018). However, it consists of a rhythmic succession of interbedded skeletal muddy limestone and shale suggesting either variable siliciclastic input into a protected normal-marine environment, or a platform was subject to influx of deeper water (shale) facies (Oruche et al., 2018).

The Millbrig bentonite is exposed in the eastern embayment, at Loc. 1 (Fig. 3.1; see coordinates in Appendix A), and dated isotopically (U-Pb in zircon) at 453.36 ± 0.38 Ma (Oruche et al., 2018). This altered volcanic ash forms the uppermost part of the L’Orignal Formation at Loc. 1, but the bentonite has not been found at other sites in the embayment suggesting that either a disconformity caps the L’Orignal Formation or depositional conditions prevented its accumulation. The bentonite is traced throughout the eastern United States (Mitchell et al., 2004; Sell et al., 2015) and is a marker of the North American Turinian-Chatfieldian stage boundary (Leslie and Bergström, 1995). Reworked altered volcanic ash of possible Millbrig affinity occurs in basal Chatfieldian strata that disconformably overlie the Selby Formation along the northern limit of the Appalachian Basin in southern Ontario (Loc. 13 in Fig. 3.1, and Fig. 3.2; Oruche et al., 2018). This event approximates timing of significant retro-arc shortening in the hinterland, with regional platform step-back along the foreland (Karabinos et al., 2017; Macdonald et al., 2017).
In the Ottawa Embayment, the Chatfieldian Rockland and overlying Hull formations were formally defined from Stewart Quarry (Loc. 3), the type section of the lowermost Trentonian succession in eastern North America (Fig. 3.2; Raymond, 1914; Kay, 1937). The Rockland-Hull succession was previously regarded to represent a conformable superposition (Barta et al., 2007). Re-examination (Oruche et al., 2018), however, suggests regional platform segmentation following Millbrig deposition, with formation of a mosaic of low-energy muddy carbonate banks (Rockland Formation) laterally adjacent to deeper-water platform environments (lower Hull Formation), previously referred to as micro-seaways (Oruche et al., 2018). Subsequent shallowing resulted in regional expansion and coalescence of high-energy crinoid shoals (middle through upper Hull Formation), with local overstepping of the Rockland carbonate banks by coarse-grained Hull facies (Oruche et al., 2018). This history is contemporary with similar shoal development in the western Quebec Basin (Mehrtens, 1988).

**Methodology and analytical protocols**

**Field methods**

Lithostratigraphic sections form a transect of ~220 km along the axis of the Ottawa Embayment and northwest into outliers within the Ottawa Bonnechere graben (Fig. 3.3). Locality numbers are maintained from Oruche et al. (2018), with additional sites (Loc. 21-23) reported here. Geographic coordinates for localities are provided in Appendix A. Details of formation lithofacies were presented by Oruche et al. (2018) and Gazdewich (2018), with a graphic section for Loc. 21 provided in Appendix D.
**Fig. 3.3:** $\delta^{13}$C stratigraphic profiles through the Lowville through Hull succession extending from the Ottawa Embayment into Ordovician outliers along the Ottawa-Bonnechere graben. Interpreted $\delta^{13}$C correlation ties are based on lithostratigraphic constraints (see text) including a datum (thick black line) represented by a sequence boundary capping the Watertown Formation, and pattern matching within the Watertown Formation. Four positive $\delta^{13}$C excursions (E1-E4) are indicated, including local high-order subdivisions (e.g., E3a-c). Other unnamed excursions are apparent in the Hull Formation. Bentonites (k1, k2) and Millbrig Bentonite (M) are indicated. The division between carbonate banks and deeper water platform settings for the Rockland – lower Hull interval is indicated.
About 580 rock samples were collected for $\delta^{13}$C and $\delta^{18}$O analyses. In general, vertical sample spacing was approximately every 50 centimetres, with additional samples taken immediately bounding formation contacts, selected hardgrounds, and disconformities. The target lithology was micrite, but this facies is largely absent from the Hull Formation, so the finest fractions of packstone and grainstone were used instead.

**Isotopic analytical protocols**

Samples were cut and cleaned, and powder was obtained by microdrilling the fresh surface avoiding fossil material and secondary void-fill carbonate. Samples from Loc. 1, 2, and 9 were processed at the Queen's Facility for Isotope Research (Kingston), whereas analyses for the remaining sites (3-7, 12, 21-23) were conducted at the Ján Veizer Stable Isotope Laboratory (University of Ottawa). Isotope data are tabulated in Appendix F.

At Queen's Facility for Isotope Research (Queen’s University, Kingston, Canada), the $\delta^{18}$O and $\delta^{13}$C ratios of calcite were determined by reacting approximately 1 mg of powdered material with 100% anhydrous phosphoric acid at 72°C for 4 hours. The CO$_2$ released was analyzed using a Thermo-Finnigan Gas Bench coupled to a Thermo-Finnigan DeltaPlus XP Continuous-Flow Isotope-Ratio Mass Spectrometer (CF-IRMS). $\delta^{18}$O and $\delta^{13}$C values are reported using the delta ($\delta$) notation in permil (‰), relative to Vienna Standard Mean Ocean Water (VSMOW, VSLAP) and Vienna Pee Dee Belemnite (VPDB), respectively. Precision was based on random duplicate analysis, and an accuracy (standard deviation) of 0.2 ‰ for both $\delta^{13}$C and $\delta^{18}$O.

Methodology for analysis at the Jan Veizer Stable Isotope Laboratory (University of Ottawa, Ottawa, Canada) is provided at [https://isotope.uottawa.ca/en/services-solids](https://isotope.uottawa.ca/en/services-solids).
summary, samples are measured into clean exetainers, acidified (0.1mL of anhydrous phosphoric acid), and the containers flushed and filled with UHP helium off-line for 4 minutes at a rate of 60-70 mL/min. Prepared vials were then placed in the heated (25.0°C for calcite) block of the GasBench and left to react for 24 hours. Measurements were performed on a Thermo-Finnigan Delta XP and a Gas Bench II, with an analytical precision of +/- 0.15 ‰ for δ¹³C and δ¹⁸O. Both isotope rates were normalized using international standards relative to VPDB: NBS-18 (δ¹³C: -5.01; δ¹⁸O: -23.00); NBS-19 (δ¹³C: 1.95; δ¹⁸O: -2.20); and, LSVEC (lithium carbonate: δ¹³C: -46.5; δ¹⁸O: -26.64), which is under review for δ¹⁸O.

Data presentation

δ¹³C profiles in Figure 3.3 are also tabulated along with corresponding δ¹⁸O values in Appendix F. Combined δ¹⁸O and δ¹³C stratigraphic profiles and cross-plots of these variables are presented in Appendix G, and allow for identification of positive or negative covariant trends, especially across stratigraphic boundaries. Covariance between stable C and O isotopes is often used as a tool to define fluid mixing and-or diagenetic influence (Allan and Mathews, 1977; Immenhauser et al., 2003; Colombié et al., 2011; Metzger and Fike, 2013; Oehlerl and Swart, 2014), but vadose diagenesis may show no relationship (Swart and Oehlerl, 2017). Selected patterns found in our sections are illustrated in a hypothetical log section (Fig. 3.4).

Positive δ¹³C excursions are defined by an anomalous increase in δ¹³C values above a pre-excursion baseline (Fig. 3.4). For comparison among localities in the embayment, and with sites outside the Ottawa Embayment (Appendix H), we used a 3-point running
Fig. 3.4: Hypothetical lithostratigraphic and chemostratigraphic profiles ($\delta^{13}$C, $\delta^{18}$O) illustrating four patterns (A-D) of stratigraphically restricted isotopic variation and excursions, 1-4. Type A is noted beneath disconformities; type B appears to modify the margins of $\delta^{13}$C excursions, thereby improving their delimitation; type C, with or without $\delta^{18}$O, represents the core of $\delta^{13}$C excursions; and, type D represents variation not otherwise linked to obvious lithostratigraphic variation. Formations a-e are hypothetical formations.
average through a given profile in order to approximate the maximum ($\delta^{13}C_{\text{max}}$) value for a given excursion relative to a pre-excursion baseline. This approach removes the influence of individual extreme values relative to lower bounding values in the profile. In one example (Metzger et al., 2014), values could not be extracted confidently from published profiles and estimates are based on visual determination only. Derived averaged profiles for sections in the embayment are shown with original data (Appendix F). From this derived database, geographic variation for baseline and maximum $\delta^{13}C$ values, and the difference ($\delta^{13}C\Delta$) between these values, is presented in a series of maps for each excursion, including the GICE.

**Results**

$\delta^{13}C$ profiles of the Lowville through Hull formation succession in the Ottawa Embayment illustrate four positive excursions (E1-E4) ranging up to $\sim 3\, \%$ relative to local baseline values (Fig. 3.3). The first three are the most confidently defined and correlated along the transect (Fig. 3.3). We describe the stratigraphic frameworks that allow for excursion correlation; demonstrate that the excursions replicate previous documentation of similar events in the embayment (Barta et al., 2007; Riopel et al., 2018); show that these excursions are not significantly affected by diagenesis; and, summarize their intrabasinal character and variation within the embayment.

**Stratigraphic frameworks for $\delta^{13}C$ correlation**

Excursions E1 and E2 are stratigraphically differentiated on the basis of disconformities that bound the Watertown Formation. Excursion E1 is restricted to this formation: it post-
Fig. 3.5: Correlation of lithostratigraphy and δ¹³C variation between the type section of the Rockland Formation (Loc. 3) and Loc. 2, ~33 km to the east. For Loc. 3, our profile is comparable with that of Barta et al. (2007). δ¹³C excursions are labelled in bold with possible higher order variation (a-c) illustrated for excursion E3, and its separation from the younger E4 across a prominent disconformity. Distribution of conodont assemblages (Barnes, 1964, 1967), biozones (Barta et al., 2007), and related sample distributions are illustrated. Abbreviations: Ch, Chaumont; tr, transitional; R, Rocklandian; L’O, L’Orignal; Pam, Pamela; Hull3, Unit H3 (upper Hull Formation); W, Watertown.
dates a regional post-Lowville disconformity that coincides with an apparent regional marker, a slight negative deflection in δ\textsuperscript{13}C (L-W in Fig. 3.3), except Loc. 2. The upper limit of the excursion is the post-Watertown disconformity that defines a basin-wide sequence boundary (Oruche et al., 2018). As the underlying Lowville Formation is regionally uniform in thickness, except at Loc. 5, and consists of peritidal to shallow subtidal carbonate facies, we interpret the post-Lowville paleosurface as having been regionally horizontal prior to differential subsidence that accommodated substantial differential thickness within the embayment (Fig. 3.3). Thickness variations and lateral changes in the magnitude of excursion E1 are described in more detail below.

The base of excursion E2 is lithostratigraphically constrained by the post-Watertown sequence boundary. At Loc. 1 in the eastern embayment, it occurs within the L’Orignal Formation and beneath the Millbrig bentonite that occurs at the top of this formation at this locality (Oruche et al., 2018). Similar, and more pronounced, excursion profiles occur within this formation at other localities (Fig. 3.3), but the Millbrig ash bed is not preserved at these other sites.

Three excursions (E2-E4) are distinguished and correlated between localities 2 and 3 (Fig. 3.5) by integrating lithostratigraphy and available conodont assemblage data (Schopf, 1966; Barnes, 1964, 1967; Barta et al., 2007). At Loc. 3, the type section of the Rockland Formation, Barta et al. (2007) documented a positive excursion from samples originally collected by Schopf (1966) for conodont analysis. The excursion appears in the lower Rockland Formation and extends up into overlying calcarenites that they referred to the Kings Falls Formation (Fig. 3.5). These authors interpreted this to be the local expression of the Guttenberg δ\textsuperscript{13}C excursion (GiCE).
A greater sampling density for this site (Fig. 3.5) reveals a higher-order $\delta^{13}$C variation that enables separation of excursion E2 within the L’Orignal Formation. A younger excursion E3 in the overlying Rockland strata (Fig. 3.5) consists of possible subordinate peaks (a-c) that can be correlated with a similar profile at Loc. 2, ~33 km to the east. At Loc. 3, however, the disconformity that separates calcarenite of the Hull (or Kings Falls) Formation from the underlying Rockland Formation also separates parts of two excursions, E3 and E4 (Fig. 3.5). Conodont data do not resolve the potential expanse of time gap represented by the disconformity, a potentially significant lacuna (Wheeler, 1964) consisting of post-E3 erosion and a period of non-deposition prior to accumulation of sediment hosting excursion E4 (Fig. 3.6). The Hull calcarenites contain the *Belodina confluens* biozone (Fig. 3.5; Barta et al., 2007) but the older *Plectodina tenuis* biozone is not recognized at Loc. 3 and Barta et al. (2007) defined the interval containing excursion E3 as a combined *P. undatus*-*P. tenuis* zone (Fig. 3.5). At Loc. 2, distribution of Barnes’ (1964, 1967) Rockland conodont assemblage illustrates a comparable association with rocks host to E3 at Loc. 3. Our interpretation is that only part of excursion E3 (or the GICE) is represented at this site, with rocks that hosted the post-peak decline to a baseline value having been eroded. Stratigraphic superposition of excursions E1 and E2 is interpreted at sites where sustained elevated $\delta^{13}$C values extend across the post-Watertown sequence boundary (Fig. 3.3).

**Validation of $\delta^{13}$C excursions**

The potential for intrastratal isotopic variation arising from primary compositional heterogeneity and diagenesis (Ludvigson et al., 1996; Metzger and Fike, 2013) requires
Fig. 3.6: Interpreted temporal succession of $\delta^{13}$C profile segments for Loc. 3 shown relative to erosional/hiatal gaps of uncertain magnitude along the erosional tops of the Lowville and Rockland formations. This illustrates absence of the upper part of excursion E3 due to erosion, and absence of the lower part of the excursion E4 due to non-deposition. Illustrated for reference is the age of the Millbrig bentonite that occurs at Loc. 1 in the uppermost L’Orignal Formation (Oruche et al., 2018); and the requirement that the E4 excursion must be the same age or younger than the age of the base of the Belodina confluens zone (from Webby et al., 2004).
substantiation of both definition and reproducibility of excursions. We provide two measures that suggest that the general pattern and magnitude of our defined excursions, if not higher order variation, are significant; and, that the overall magnitude has not been impacted by post-depositional diagenesis.

The combined E2-to-E4 stratigraphy at Loc. 3 illustrates a net shift above baseline values and a 1st-order pattern similar to the original excursion defined by Barta et al. (2007) at this same locality (Fig. 3.5). Confirmation of our higher order variation is not possible due to differences in sampling density. Within the Watertown and Rockland formations, however, our δ¹³C values are up to 0.8 ‰ lower in lime mudstone whereas the two datasets are indistinguishable in gravel-sized grainstones in both the Rockland and Hull formations (Fig. 3.5). The reason for such a difference remains uncertain but because Barta et al. (2007) used rock samples originally collected for conodont analysis (Schopf, 1966) there may be some gravel-sized bias inherent in samples selected for conodont processing.

Our second validation measure relates to the potential impact of diagenesis on stratigraphic form and magnitude of a given δ¹³C profile. Modification of the carbon isotopic system during burial diagenesis in shallow-water carbonate successions is usually viewed to be conservative (Banner and Hanson, 1990; Ripperdan, 2001). However, near-surface meteoric diagenesis during exposure, potentially influencing centimetres to tens of metres of the underlying stratigraphic section, may or may not produce a covariant negative shift in C and O isotope ratios (Allan and Matthews, 1977; Oehlert and Swart, 2014; Swart and Oehlert, 2017). Measured statistical co-dependence
using coefficient of determination ($r^2$) can provide the basis for interpreting the influence of diagenesis (Metzger and Fike, 2013; Metzger et al., 2014). Our dataset (see Appendix G) reveals the majority of $r^2$ values to be $< 0.20$, with the highest ($< 0.40$, Loc. 23) still representing a weak relationship. For comparison, values of greater than $\approx 0.6$ (Metzger and Fike, 2013), and $\approx 0.3$ to $\approx 0.7$ (Metzger et al., 2014) are considered significant. However, isotopic covariance over a narrow stratigraphic interval, such as beneath a disconformity, may not possess sufficient statistical impact compared to a record of no covariance in the rest of a section. Instead, this relationship is revealed by examining δ$^{13}$C- δ$^{18}$O stratigraphic profiles (see Appendix G), and we recognize four patterns (A-D; Fig. 3.4) of stratigraphically restricted isotopic variation within our dataset:

(1) Type A is represented by a negative shift in both isotope values beneath an erosional paleosurface (Fig. 3.4). This is developed over tens of centimetres to metres in some cases within the uppermost Lowville Formation (e.g. Locs. 2, 9, 22, 23) and Watertown Formation (Locs. 5, 7, 23) (see Fig. 3.4 and Appendix G) beneath regional disconformities that cap each formation.

(2) Type B represents covariant negative shifts in both isotope ratios that may enhance or more sharply delimit the pre- and post-maximum gradients (Fig. 3.4).

(3) Type C characterizes our defined positive excursions, E1 to E4, some with covariant positive shifts in δ$^{18}$O (Fig. 3.4). Overall, this pattern occupies much thicker stratigraphic intervals (metres to ten’s of metres) than the other types of covariant isotopic variation, best demonstrated by excursion E1 in the Watertown Formation (Fig. 3.3).
Type D reflects variation in one isotope system but not the other (Fig. 3.4) and occurs randomly through sections.

**Description of δ\(^{13}\)C excursion patterns**

*Excursion E1*

This excursion displays a well-developed higher order variation (E1a to e) best defined and correlative within the central part of the embayment (Fig. 3.3). Across the embayment, but absent at Loc. 7, a negative deflection (< 1 ‰) marks the excursion base. Along the central part of the transect, or toward the basin center, the excursion is defined by a prominent upsection shift (up to 2.8 ‰) of relatively sustained values relative to section baselines of ~ 0 ‰. In the eastern and western sectors of the embayment, where the formation is thinner, the excursion contains pronounced (< 2 ‰) variation over short (1-2 m) stratigraphic intervals (Fig. 3.3).

The interpreted intraformational correlation of subordinate variation (E1a to e) suggests that the lower parts of the Watertown Formation are not represented to the west or east but onlap paleotopography developed on the Lowville Formation (Fig. 3.3). This is supported by transgressive deposits appear to the west (Loc. 6 and 9) and less well developed to the north (Loc. 3). Moving to the east and west, the higher order correlation becomes speculative given potential section cut-out, in addition to evidence for increased stratigraphic condensation moving into the western embayment (e.g., Loc. 9; Oruche et al., 2018). Correlation of this subordinate variation does not show appreciable thinning away from the center of the embayment. However, the cross-sectional geometry of
excursion E1e extending toward Loc. 9 suggests significant (metre-scale) truncation of
the excursion and host formation at the post-Watertown Formation sequence boundary.

A covariant decrease in δ\textsuperscript{13}C and δ\textsuperscript{18}O occurs in uppermost Watertown strata beneath the
sequence boundary. At Locs 1, 5, 6, and 21 (Fig. 3.3), an upsection decline in δ\textsuperscript{13}C values
begins metres below this uppermost change at other sites. It is uncertain if this relates to
diagenetic influence or reflects post-peak decline in excursion E1.

Subordinate maxima of excursion E1 coincide with lithofacies succession within the
Watertown Formation. At Loc. 9, the two peaks comprising excursion E1d are associated
with, respectively, basal coarse-grained transgressive deposits developed across the
eroded Lowville paleoplatform and thick-bedded muddy carbonate facies that otherwise
characterizes a regional expression of this formation (Oruche et al., 2018).

Geographic variation in pre-excursion baseline values, δ\textsuperscript{13}C\textsubscript{max} (based on 3-point running
average; Appendix F) and the computed difference of these endmember values, or δ\textsuperscript{13}C\textsubscript{Δ},
are illustrated (Fig. 3.7) for sections in the embayment and outliers (Locs. 22 and 23) in
the graben extension to the northwest (Fig. 3.3). The western embayment (Locs. 6 and 9)
contains the lowest baseline values, there being a well constrained abrupt increase across
the trace of the present-day Gloucester Fault between Locs. 6 and 7 (Fig. 3.7A). Sites
along the east side of the fault trace display a common range in values, with subsequent
increase farther east to the three remaining sites as noted west of the western embayment
(Fig. 3.7A). δ\textsuperscript{13}C\textsubscript{max} variation suggests a similar abrupt increased in values along the fault
trace (Fig. 3.7B), also well illustrated in the regional cross-section (Fig. 3.3), traced over
Fig. 3.7: Excursion E1 geographic distribution of pre-excursion baseline and maximum δ¹³C values, and computed difference (δ¹³CΔ), hosted by the Watertown Formation: (A) Pre-peak baseline values; (B) δ¹³Cmax values; and (C) δ¹³CΔ values. Regional context of locality distribution can be compared with Fig. 3.3. Present-day fault traces are shown in light grey: GF, Gloucester Fault; RF, Rideau Fault; dashed lines, boundary.
to the far eastern limit of our map region. Regional variation in the difference (or $\delta^{13}C_a$) in baseline and $\delta^{13}C_{max}$ values creates a regional blocky mosaic with evidence that the boundaries separating value ranges are geographically abrupt as illustrated between Locs. 6 and 7, and between Locs. 1 and 2.

Excursion E2

Restricted to the L’Orignal Formation, the vertical extent of this excursion varies with formation thickness. Excursion E2 is defined by either a prominent positive shift (< 2-3 ‰) across the Watertown-L’Orignal formation boundary, and above a negative deflection within the uppermost Watertown Formation; or, as in the case of Locs. 2 and 21, sustained elevated values are carried across the formation boundary from E1 reflecting excursion superposition similar to that described for excursions E3 and E4 at Loc. 3 (Fig. 3.5).

A comparison of conodont assemblage data for Locs. 2 and 3 suggests possible diachroneity along the base of the L’Orignal Formation (Fig. 3.5). This formation represents the lower few metres of previous workers’ Rockland Formation. At Loc. 2, the L’Orignal Formation straddles Barnes’ (1964, 1967) assemblage of Chaumontian (Watertown Formation-hosted) fauna and an assemblage marking a transition into a Rocklandian (early Chatfieldian) assemblage (Fig. 3.5). At Loc. 3, the L’Orignal Formation lies within Barnes’ (1964, 1967) transition from Turinian or Chaumontian fauna and a Rocklandian assemblage (Fig. 3.5). Regional lithostratigraphic analysis (Oruche et al., 2018) suggests a regional mosaic of initial low to high-energy basal carbonate facies, with Loc. 3 host to an initial high-energy paleoenvironment. Combined
with conodont data, this may characterize a subtle diachronicity of onset of deposition with high-energy sites (Loc. 3) forming slightly shallower paleobathymetry, transgressed later than adjoining regions (Loc. 2).

Geographic variation in $\delta^{13}C_{\text{baseline}}$ and $\delta^{13}C_{\text{max}}$ values (Fig. 3.8A) shows low values associated again with the western embayment, and an abrupt increase across the trace of the Gloucester Fault (Fig. 3.8A). Indeed, there appears to be a repetition of an east-directed decrease in $\delta^{13}C_{\text{max}}$ values from, first, the outliers (Locs. 22 and 23) into the western embayment (Locs. 6 and 9), then extending east of the trace of the Gloucester Fault (Fig. 3.8B). The $\delta^{13}C_{\Delta}$ map reveals a common range of values east of the Gloucester Fault apart from Loc. 3, compared to an eastern decrease into the western embayment (Fig. 3.8C).

**Excursion E3**

Excursion 3, or the local expression of the GICE (Fig. 3.5), post-dates regional platform segmentation following deposition of the Millbrig ash fall, and is coeval with a regional mosaic of deeper water platform environments of shale and skeletal-rich limestone that form unit H1 of the Hull Formation, and quiet-water muddy carbonate banks characterizing facies of the Rockland Formation (Fig. 3.3; Oruche et al., 2018). The excursion occurs in both formations (Fig. 3.3) but definition of excursion E3 in the thicker successions of the lower Hull Formation at Locs. 4, 5, and 21 remains uncertain because of the absence of regional lithostratigraphic and biostratigraphic constraints. In the western embayment, an overall thinner Rockland-Hull succession contains more
Fig. 3.8: Excursion E2 geographic distribution of pre-excursion baseline and maximum $\delta^{13}C$ values, and computed difference ($\delta^{13}C_{\Delta}$), hosted by the L’Orignal Formation: (A) Pre-peak baseline values; (B) $\delta^{13}C_{\text{max}}$ values; and, (C) $\delta^{13}C_{\Delta}$ values. Regional context of locality distribution can be compared with Fig. 3.3. Present-day fault traces are shown in light grey: GF, Gloucester Fault; RF, Rideau Fault; dashed lines, boundary.
pronounced higher order intersample variation in δ¹³C values compared to the interpreted thicker expressions of this excursion in the embayment center (Fig. 3.3). A similar, but subdued high-order variation profile pattern re-appears in the easternmost part of the embayment, at Loc. 1 (Fig. 3.3).

At the time of thesis submission, the GICE had been reported from the Hull Formation at Loc. 6 (Riopel et al., 2018) hosted by a thick succession of lower (H1) shale-limestone interbeds, with only the lower few metres of this interval represented in our profile (Fig. 3.3). Our profile shows elevated δ¹³C values across the disconformable L’Orignal-Hull formation boundary (Fig. 3.3), and illustrates superposition of two excursions (Fig. 3.5).

Definition of geographic variation in δ¹³Cbaseline values for excursion E3 is incomplete due to our inability to define these endmember values because of superposition of excursions or, maybe, due to a too coarse a sampling interval; and, uncertainty about correct identification of the position of E3 at some sites (Fig. 3.3). Available baseline data, however, suggest a possible regional decrease from west to east (Fig. 3.9A). As with excursion E2, δ¹³Cmax values illustrate a well constrained repetition of an east-directed decrease in values across the map area, with the trace of the Gloucester Fault forming the apparent boundary (Fig. 3.9B). However, there is abrupt variation moving south from Loc. 21 to Loc. 4, also extending across the trace of the Rideau Fault. With regard to δ¹³CΔ variation, our dataset suggests that elevated values are affiliated with carbonate bank strata represented by the Rockland Formation (Loc. 3 and 2) compared to strata within the adjacent microseaway facies (Loc. 21 and 1).
**Fig. 3.9:** Excursion E3 or the GI CE geographic distribution of pre-excursion baseline and maximum $\delta^{13}C$ values, and computed difference ($\delta^{13}C_\Delta$), positioned in the Rockland and lower Hull formations: (A) Pre-peak baseline values; (B) $\delta^{13}C_{\text{max}}$ values; and, (C) $\delta^{13}C_\Delta$ values. Regional context of locality distribution can be compared with Fig. 3.3. Present-day fault traces are shown in light grey: GF, Gloucester Fault; RF, Rideau Fault; dashed lines, boundary.
Excursion E4

This excursion is hosted by the upper (unit H3) Hull Formation (Fig. 3.3). Although defined at Loc. 3, we are less confident about its stratigraphic position and correlation among sites due to both absence of obvious lithostratigraphic marker beds and differential preservation of the upper Hull succession across the region. The interpreted absence of the pre-peak increase in $\delta^{13}C$ for this excursion at Loc. 3 (Fig. 3.6) contrasts with other sites (Locs. 1, 9, 21 and 23) that display the increase demonstrating a complete excursion. Geographic variation in $\delta^{13}C_{max}$ may bracket $\sim$2 ‰, but there are too few sections host to this excursion, and insufficient correlation constraints, to warrant confidence in defining additional patterns.

Discussion

Intrabasin correlation of Turinian $\delta^{13}C$ excursions in the Ottawa Embayment is tightly constrained on the basis of correlation among closely-spaced sections of paleoerosional surfaces bounding the Watertown Formation, and the appearance of the Millbrig bentonite capping the L’Orignal Formation at Loc. 1 (Fig. 3.3). A similar approach is more difficult for the Chatfieldian section because of the absence of lithostratigraphic markers passing through coeval but environmentally distinct platform facies. Nonetheless, our study resolves prominent lateral variation in magnitude of carbon-isotope excursions accompanying variation in formation thickness and facies, and the problematic superposition of excursions across disconformities. We discuss the apparent roles of intrabasinal oceanography, depositional environments, and syndepositional tectonism modulating the record of Late Ordovician $\delta^{13}C$ excursions.
Turinian-Chatfieldian excursions in Southern Laurentia

The Turinian $\delta^{13}$C excursions, E1 and E2, and the Chatfieldian GICE in the Ottawa Embayment correlate with coeval isotopic events through the northern Appalachian Basin (Ontario) and further afield in the Mississippi Valley of the mid-continental United States (Fig. 3.10). The younger excursion E4, when integrated with improved biostratigraphic information, may resolve high-order inter-regional correlation within the Chatfieldian succession.

New observations from our study related to the regional continuity of $\delta^{13}$C excursions (Fig. 3.10) are summarized as follows. First, within the Turinian succession, the Quimby’s Mill Formation (Iowa) appears to be equivalent to the Lowville Formation of New York (Fig. 3.2) such that the same-named excursion and the Grand Detour Excursion may be correlative with subdued excursions in the Lowville of the Ottawa Embayment as found at Locs. 3 and 23. Second, excursion E1 hosted by the Watertown Formation does not appear to have a correlative record in the mid-continent region likely due to the effect of late Turinian stratigraphic condensation, non-deposition, and-or erosion in this region (Ludvigson et al., 2004). However, it is traced into the northern Appalachian Basin (Ontario) suggesting that it may extend farther south through equivalent strata in the eastern United States (Brett et al., 2004). Third, the Specht’s Ferry Excursion (SFE) in mid-continent USA may correlate with excursion E2 in the Ottawa Embayment and a coeval excursion in the Selby Formation at Loc. 13 along the northern limit of the northern Appalachian Basin (Ontario). Fourth, with regard to the GICE, our correlation suggests that a previously noted excursion in the Napanee Formation at the Dexter Quarry (New York), which had been interpreted as a local expression of the GICE
Fig. 3.10: Proposed regional correlation of δ¹³C excursions (E1-E4) from the Ottawa Embayment into the northern Appalachian Basin (Ontario, New York) and mid-continent USA (Iowa). The E3 excursion is interpreted to be equivalent to the Guttenberg δ¹³C excursion (GICE). The uppermost excursion in the Dexter Quarry profile may be equivalent to excursion E4, not the GICE as previously interpreted (Barta et al., 2007; Bergström et al., 2010b). Abbreviations: SFE, Spechts Ferry excursion; QME, Quimbys Mill excursion; GDE, Grand Detour excursion; C, Carimona Formation; G, Glencoe Formation.
(Barta et al., 2007), is probably younger, maybe associated with excursion E4. Diagnostic conodont-zone species have yet to be found in this section, and the excursion lies well above (~10 m) the position of the Millbrig bentonite (Mitchell et al., 2004). Regionally, the loss of section between excursions E3 and a putative E4 at Dexter Quarry moving north toward Loc. 3 in the embayment could be explained by regional downlap of high energy calcarenites (Hull Formation or King Falls Formation) across the eroded carbonate banks characterizing the Rockland succession. Erosional truncation of the GICE at Loc. 3 (Fig. 3.5) demonstrates that care must be taken to ensure separation of distinct excursions when establishing regional correlation. The time span characterized by the disconformity at Loc. 3 is uncertain but might be resolved in the future by looking for any offset in conodont or bulk-rock Sr-isotope ratios across the disconformity; there was a relatively rapid change in the Sr-isotope ratio of seawater at this time in the Ordovician (Saltzman et al., 2014; Edwards et al., 2015).

**Turinian geographic variation in δ13C (Ottawa Embayment)**

Our δ13C_{carb} dataset, when integrated with distribution and character of lithofacies, and formation thickness, provides a basis to begin resolving the many influences controlling spatial and temporal variation in 13C enrichment and depletion (Patterson and Walter, 1994; Ripperdan, 2001; Metzger and Fike, 2013). We recognize that additional datasets including C_{org} and, in particular, its biomarker sources (Pancost et al., 2006), are important to constrain our interpretations but these are not available at this time.

Two prominent patterns emerge with regard to geographic variation in Turinian δ13C excursions: 1) their stratigraphic restriction to formations and specific facies types; and 2)
abrupt changes (enrichment, depletion) spatially associated with the trace of present-day regional faults. Excursions E1 and E2 in successive formations are each associated with a period of marine transgression following carbonate-platform erosion. This does not, necessarily, define a causative association, but may only indicate that a sedimentary host was available to capture a regional or global signal unrelated to transgression. However, we note that the stratigraphy of $^{13}$C enrichment is associated with basal high-energy transgressive deposits or within the succeeding lower energy muddy carbonate facies or in both, as illustrated by peaks of subordinant E1d excursion at Loc. 9. This suggests a stratigraphic pattern that may be related to local controls that, in turn, modulate a regional signal. If correct, transgression allowed for preferential $^{12}$C drawdown in response to elevated productivity (Ripperdan, 2001) during renewal of marine conditions across a previously eroded platform; then, renewed sequestration of $^{12}$C within the succeeding muddy carbonate facies may be associated specifically with burial of fines, including organics (Ripperdan, 2001), associated with quieter, deeper water deposition.

Positive shifts in $\delta^{13}$C excursions related to transgressive episodes have also been related to changing salinity as transgression mixes marine and existing meteoric waters associated with prior exposure of a platform (Immenhauser et al., 2003). Local covariant negative shifts in $\delta^{13}$C and $\delta^{18}$O occur immediately beneath the Lowville and post-Watertown disconformities suggesting meteoric alteration (Allan and Matthews, 1977; Oehlert and Swart, 2014). However, the required positive C and O isotopic covariance (Immenhauser et al., 2003) is difficult to demonstrate in our sections, possibly due in part to erosional truncation across the platform, or it occupies an interval thickness below our sampling resolution. Superimposed on the above regional stratigraphy of excursions is a
pattern of geographic variation in $\delta^{13}\text{C}_{\text{carb}}$ per excursion (Fig. 3.7 and 3.8). As noted, our isotope datasets (Appendix G) do not indicate obvious significant formation-scale diagenetic influence that can explain such regional variation. Instead, we interpret the action of syndepositional structural control that resulted in locally changed oceanographic and sedimentary patterns.

For excursion E1, depletion in $^{13}\text{C}$ in the western embayment (Loc. 6 and 9) is associated with evidence of stratigraphic condensation defined not only by westward-thinning of the Watertown Formation but also abrupt appearance and abundance of hardgrounds. These omission surfaces were likely formed through elevated wave erosion during short-lived lowering of sea level (Oruche et al., 2018). East of the trace of the Gloucester Fault, formation thickness increases, omission surfaces disappear, and there is a coincident $^{13}\text{C}$ enrichment (Fig. 3.7 and 3.8). We interpret this lateral contrast to a subtle but significant eastward deepening across the trace of the present-day fault wherein the western embayment was slightly elevated and the area east of the fault trace formed a sub-basin subject to elevated subsidence and accumulation rates. This is reflected, lithostratigraphically, by west-directed appearance of higher energy transgressive deposits across the underlying eroded Lowville platform in the western embayment and moving north in the region of Loc. 3. Stratigraphic and paleostress analyses have demonstrated that an ancestral Gloucester Fault was active episodically during the Middle and Late Ordovician (Salad Hersi and Dix, 1999; Rimando and Benn, 2005; Dix and Al Rodhan, 2006; Dix and Jolicoeur, 2011; Oruche et al., 2018).

An important effect of such structural geometry (Fig. 3.11) is increased oceanographic restriction of seawater exchange across the structural margin, wherein reduced exchange
Fig. 3.11: Interpreted platform stratigraphy within the Ottawa Embayment and adjacent graben during the late Turinian and Chatfieldian, and lateral patterns in $^{13}\text{C}$ enrichment and depletion relative to oceanographic parameters. Thin black lines represent stratigraphic fabric, highlighting preferential thickening (wider spacing) versus stratigraphic condensation (thinly spaced). Vertical dashed lines define interpreted locations of structural (fault) control. The cross-sections are placed in regional context with regard to platform development in the western Quebec Basin (Harland and Pickerill, 1982) and structural events in the orogen (Karabinos et al., 2017; Macdonald et al., 2017). Abbreviations: SL, sea level; GF, Gloucester Fault.
late Turonian

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<th>graben</th>
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<th>western Quebec Basin</th>
<th>proximal foreland-orogen</th>
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<tbody>
<tr>
<td>Locs. 22 and 23</td>
<td>Locs. 6 and 9 and 21</td>
<td>Loc. 1</td>
<td>subduction polarity reversal</td>
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E1: $^{13}C$ ↓ \[ ^{13}C \downarrow \]
E2: $^{13}C$ ↑ \[ ^{13}C \uparrow \]

(1) carbonate bank: low energy
(2) carbonate bank: low energy, with periods of wave erosion; stratigraphic condensation; hardgrounds, omission surfaces
(3) carbonate basin: low energy; stratigraphic condensation; hardgrounds

GF: elevated accumulation and subsidence rates

Chatfieldian: post-453 Ma

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<th>Ottawa Embayment</th>
<th>western Quebec Basin</th>
<th>proximal foreland</th>
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<tr>
<td>Locs. 22 and 23</td>
<td>Locs. 6 and 9</td>
<td>Loc. 3</td>
<td>retro-arc shortening</td>
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<td>Loc. 1</td>
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E3: $^{13}C$ ↑ \[ ^{13}C \uparrow \]

(1) carbonate bank: low energy; mostly interbedded lime mudstone and minor shale
(2) deeper water platform: high-to-low energy; skeletal rudstone interbedded with shale

GF: heterotrophic productivity

photosynthetic drawdown (?)

SL: diachronous platform collapse

restricted seawater exchange
high-energy currents

shale
platform carbonate, with shale
orogen-derived sand

oceanographic parameters
with an open-ocean reservoir leads to greater residence time and depletion in $^{13}$C on the adjacent elevated platform (Patterson and Walter, 1994). Wave erosion on the elevated portion of the Watertown platform would enhance greater recycling and oxidation of $C_{org}$, and this may explain some of the high-order variation of excursion E1 in this region (Fig. 3.3). Conversely, east of the fault trace, a sub-basin is formed, one that exhibits increased $^{13}$C enrichment toward the center (Fig. 3.7B). East of the embayment the equivalent Turinian succession in the Montreal region (Fig. 3.1) is only 4-7 m, indicating significant thinning (~50%) into the western Quebec Basin. This may indicate that the central to eastern embayment formed a sub-basin, with relative uplift to the east, resulting in a restricted water mass relative to open-ocean exchange. A similar structural control has been proposed to explain regional differential thinning of the Pamelia Formation (Harland and Pickerill, 1982; Salad Hersi and Dix, 1999). In the created restricted basin, removal of $^{12}$C due to organic burial could influence local basin-water chemistry (Ripperdan, 2001). Coeval enrichment in $^{13}C_{max}$ within the outliers west of the western embayment (Fig. 3.7B) may also identify a slightly deeper, and restricted basin west of the area of stratigraphic condensation. A similar but less well constrained hypothesis is suggested for geographic variation in $\delta^{13}$C for excursion E2 hosted by the L’Orignal Formation. Evidence for possible depositional diachroneity across the post-Watertown sequence boundary (Fig. 3.5) and the mosaic of low to high-energy facies forming the initial stage of accumulation suggests there existed an irregular (shallow to deeper) regional bathymetry. Variable oceanographic exchange and differential productivity and accumulation rates could be established across this regional mosaic of bathymetry. Note that west of the Gloucester Fault, depletion in $^{13}C_{max}$ extends through the western
embayment into the outliers, and illustrates that the previous restricted basin in the outlier area was no longer significant, and the entire area west of the Gloucester Fault was subject to preferential recycling of C$_{org}$.

**Chatfieldian geographic variation in δ$^{13}$C (Ottawa Embayment)**

By Chatfieldian time, platform segmentation had fully established deeper platform areas and shallower low-energy carbonate banks (Oruche et al., 2018). Excursion E3 occurs in both lithofacies succession, and corresponds to the Guttenberg δ$^{13}$C excursion (GICE); a relatively short-lived (< 500,000 yr.; Ludvigson et al., 1996) and likely global event (Bergström et al., 2010b). Some workers have proposed that regional variation in δ$^{13}$C at sites across the eastern and central United States reflect regional gradients in oceanographic parameters (Young et al., 2005; Panchuk et al., 2006). Alternative models include: (1) Diagenetic alteration may have been of greater significance in creating such variation and, as a result, a regional (if not global) signal requires a driver of similar scale (Metzger and Fike, 2013); and, (2) within the current resolution of conodont biozonation, this excursion might consist of near-simultaneous events among sedimentary basins in response to eustatic sea-level rise.

In the Ottawa Embayment, the stratigraphic profile and magnitude of this event varies across the basin (Fig. 3.3 and 3.9). As with Turinian excursions, geographic variation in δ$^{13}$C$_{carb}$ illustrates some spatial relationship to the trace of the Gloucester Fault (Fig. 3.9). However, there is a stronger relationship between δ$^{13}$C magnitude and the geographic limits of the shallow bank-deeper platform boundary (Fig. 3.9). Placement of this boundary is poorly constrained in the western embayment where we have arbitrarily
placed it at the western limit of the embayment but it could occur anywhere between Loc. 9 and the northwestern outliers (Fig. 3.9). In the eastern embayment, the boundary position is well constrained between Locs. 7 and 21, and Locs. 1 and 2 (Fig. 3.9). Our isotope database (Appendix G) suggests little apparent diagenetic influence on the magnitude of this excursion across the embayment. Thus, we interpret the abrupt lateral variation to document influence of syndepositional controls (Fig. 3.11).

Platform segmentation follows the Millbrig ash fall and may represent a far-field event timed with onset of retro-arc shortening in the hinterland (Karabinos et al., 2017; Macdonald et al., 2017; Oruche et al., 2018). This local sedimentary expression coincides with regional changes to cooler and deeper water facies beyond the embayment limits, and along strike in coeval foreland basins (Ettensohn, 2008; Lavoie, 2008). For the embayment, this segmentation resulted in bathymetric differences such that there was, again, interpreted potential restriction in bank-top waters versus the higher energy deeper platforms (Fig. 3.11). The latter document fluctuating high and low energy periods of sediment accumulation (Oruche et al., 2018). Variation in sea level may also have been important given a rhythmic stratigraphy of thin shale layers that extended across the Rockland carbonate banks, smothering carbonate production. Elevated $\delta^{13}C_A$ and $\delta^{13}C_{max}$ values are restricted to the muddy platform facies of the Rockland Formation compared to the coeval skeletal-limestone and shale facies of the lower Hull Formation (Fig. 3.9B, C). Unlike the Watertown scenario of stratigraphic condensation, however, the muddy Rockland limestone facies characterize bank-top $^{13}C$ enrichment (Fig. 3.9B). This contrast may demonstrate preferential photosynthetic drawdown of $^{12}C$ (Patterson and
Walter, 1994) compared to the heterotrophic dominated productivity in the deeper water platform setting.

Variation in the GICE through Southern Laurentia

A compilation of selected published values of the GICE from drillcore and outcrop at sites across south-central Canada and middle to eastern United States (Fig. 3.12; Appendix H) reveals regional variation in magnitude utilizing a similar approach to determine baseline and maximum values, and the resulting difference, as for the Ottawa Embayment. Our compilation includes an apparent post-peak record from the northern Ottawa-Bonnechère graben (Kang, 2018) and an uncertain magnitude of the GICE and baseline from drillcore from Anticosti Island, Quebec (Fig. 3.12).

Variation in $\delta^{13}C_{\text{baseline}}$ (Fig. 3.12A) suggests that sites in mid-continent USA form a regional anomaly of lower values. If this is related to a unique aquafacies (Young et al., 2005; Panchuk et al., 2006), an absence of any significant increase toward the Appalachian orogen would appear to discount any relationship with other defined aquafacies (Fig. 3.12). If lower $\delta^{13}C$ values are related to diagenetic alteration, as proposed by Metzger and Fike (2013), then this suggests that the stratigraphic interval host to the GICE in the midcontinent has been more susceptible to alteration than elsewhere. There is no obvious regional gradient in $^{13}C_{\text{max}}$ values for the GICE (Fig. 3.12B) unless the three sites with pronounced $^{13}C$ enrichment proximal to the structural margin of the orogen define the western limit of some environmental gradient that once extended into deeper water along the Laurentian margin. The record of this would be
Fig. 3.12: Regional compilation of $\delta^{13}\text{C}$ values (baseline, maximum, and computed difference) for the GICE across southern Laurentia following ~453 Ma, the age of the Millbrig Bentonite (Sell et al., 2013; Oruche et al., 2018). Data and sources for localities outside of the study area are tabulated in Appendix H. Maximum values were calculated based on a 3-point running average as was done for sites in the Ottawa Embayment. Aquafacies of Panchuk et al. (2006) are illustrated for reference: I, Midcontinent; II, Arch Margin; III, Taconic; and IV, Southern. Paleogeography is based on Long and Copper (1987), Sanford (1993), Malo (2004), and Scotese (2014).
preserved in thrust sheets of the orogen. The map of $\delta^{13}$C also shows no obvious regional gradient (Fig. 3.12C).

At the scale of our regional compilation (Fig. 3.12), the apparent absence of any regional gradient belies local spatial variation in the Ottawa Embayment (Figs. 3.7-9) interpreted to be linked to syndepositional structural control on oceanography and sediment accumulation. These relationships might be unique to the embayment, a region of far-field structural events arising through reactivated inherited structure, and responding to distal orogenic activity in the late Turinian and early Chatfieldian (Fig. 3.11). Nonetheless, our dataset suggests that these controls modulated regional isotopic events.

**Conclusions**

Four positive $\delta^{13}$C excursions (E1 to E4) in upper Turinian-lower Chatfieldian strata are correlated with varying degrees of confidence along the axis of the Ottawa Embayment (central Canada). Excursions E1 and E2 are confined to the Turinian Watertown and L’Orignal formations, respectively; the latter coeval with the Selby Formation in the northern Appalachian Basin. Excursions E3 and E4 occur in lower Chatfieldian strata following carbonate-platform segmentation following the regional Millbrig volcanic event. Excursion E3 is correlated through coeval facies of low-energy carbonate banks (Rockland Formation) and a deeper water platform settings (Hull Formation), and forms the local expression of the Guttenberg $\delta^{13}$C anomaly. Excursion E4 occurs in younger Hull strata, its distribution poorly constrained. In the type section of the Rockland Formation, however, erosional truncation of the GICE resulted in direct superposition of excursion E4, the disconformity representing an uncertain time gap.
Turinian excursions E1 and E2 coincide with periods of transgression across eroded paleoplatforms in the embayment. Covariant negative shifts in $\delta^{13}$C and $\delta^{18}$O occur beneath the disconformities and suggest the role of diagenesis locally modifying excursion profiles. Prominent lateral variation in pre-excursion baseline values and magnitudes of both excursions are spatially associated with the trace of the present-day Gloucester Fault. We interpret syndepositional structural control having established variation in bathymetry and restriction of seawater exchange across bank-deeper platform boundaries. The shallow-water banks were subject to periods of marine erosion and register depletion in $^{13}$C likely related to enhanced recycling of $C_{org}$ and longer seawater residence time. Adjacent deeper water areas also likely represent an oceanographic restricted setting, but where elevated rates of accumulation rates and burial of $C_{org}$ resulted in $^{13}$C enrichment in the seawater.

Expression of the Chatfieldian GICE in the Ottawa Embayment reveals an opposing pattern: $^{13}$C enrichment is associated with low-energy muddy bank-top facies compared to deeper water platform facies. This relationship also requires bank-top restriction of seawater exchange, with photosynthetic drawdown of $^{12}$C likely occurring preferentially relative to conditions in the heterotrophic-dominated deeper water platform environment.

The record of late Turinian and early Chatfieldian $\delta^{13}$C excursions in the Ottawa Embayment reflect local modulation of regional, if not global, events through influence of syndepositional structural control on oceanographic exchange, water depth, accumulation rates, and productivity.
CHAPTER 4: SEQUENCE STRATIGRAPHY OF A MIDDLE TO UPPER ORDOVICIAN FORELAND SUCCESSION (OTTAWA EMBAYMENT, CENTRAL CANADA): EVIDENCE OF TECTONISM CONTROL ON SEQUENCE ARCHITECTURE ALONG SOUTHERN LAURENTIA

Submitted as:
Oruche, N. E., and Dix, G. R., Sequence Stratigraphy of a Middle to Upper Ordovician Foreland Succession (Ottawa Embayment, central Canada): Evidence for Tectonic Control on Sequence Architecture along Southern Laurentia. Submitted in Basin Research.
Abstract

The Middle to Upper Ordovician foreland succession that underlies the Ottawa Embayment in central Canada is divided into nine transgressive-regressive depositional sequences. This stratigraphy defines net deepening of the platform succession from peritidal to outer ramp settings spanning ~15 my followed by more rapid (~3 my) shallowing of a foredeep by orogen-derived siliciclastics following platform foundering. Whereas eustasy through the Middle to Late Ordovician ensured a long-term background presence of marine conditions, the following 1st-order stages of foreland-platform development appear to reflect greater far-field responses to regional tectonic controls when compared to improved geochronology of tectonic events within the orogen in the New England region and global climate change during the Ordovician. This includes: (1) onset of foreland development with Middle Ordovician closure of the Iapetus Ocean basin; (2) voluminous influx of Middle Ordovician siliciclastics related to subduction reversal and corresponding hinterland uplift, as well as global climate change to more humid conditions; (3) rapid platform deepening by the late Turinian coinciding with onset of retro-arc shortening; and (4) steepening and foundering of the platform coupled with local structural deformation and erosion during the Edenian coincided with thrust-loading along the orogen. Comparison of the nine sequences in the Ottawa Embayment to published frameworks for the adjacent Appalachian Basin that have emphasized eustatic control, reveals prominent interbasin differences in the character and position of foreland-platform sequence boundaries. This suggests that short-term intrabasinal structural controls are significant, a conclusion supported by greater compatibility with the
sequence framework of Joy et al. (2000) that emphasized tectonic influence disrupting eustatic signatures in the northern Appalachian Basin.

**Introduction**

An improved understanding of the geological framework of the Middle to Upper Ordovician foreland succession peripheral to the Appalachian orogen of eastern North America has developed through integration of litho-, bio- and chemostratigraphies ($\delta^{13}$C, Sr-isotope) coupled with age and stratigraphy of volcanic ash beds (Cameron and Mangion, 1977; Holland and Patzkowsky, 1996, 2008; Mitchell et al., 2004; Brett et al., 2004; Ettensohn, 2008; Lavoie, 2008; Bergström et al., 2010; Edwards et al., 2015; Sell et al., 2015). Likewise, an improved geochronology of orogen development (Karabinos et al., 2017) now allows definition of some foreland events as far-field responses to plate-margin evolution (Macdonald et al., 2017). There remains conflicted sequence stratigraphic frameworks (based on the sequence template of Holland and Patzkowsky, 1996, 2008) within the Appalachian Basin that underlies the eastern United States and part of southern Ontario (Fig. 4.1A) with regard to the dominance of eustasy (Brett et al., 2004; Sell et al., 2015) versus tectonism (Joy et al., 2000) in controlling the architecture and distribution of depositional systems within the foreland. The equivalent succession that underlies the Ottawa Embayment of central-eastern North America (Fig. 4.1B) provides an opportunity to test the influence of these controls by exploring concordant and discordant inter-basin correlation and nature of sequence boundaries with those of the Appalachian Basin. A depositional sequence framework is presented for the first time for the Ottawa Embayment. Incorporating timing of orogenic events (Macdonald et al., 2017), we resolve scales of tectonic influence on stratigraphic patterns within the
**Fig. 4.1:** (A) Paleogeography of southern Laurentia at ~453 Ma, along with distribution of Middle to Upper Ordovician outcrop belts (shaded grey) and approximate locations of sites incorporated within published sequence stratigraphy frameworks for the Appalachian Basin and equivalent sections in central USA (Brett et al., 2004; Sell et al., 2015; Holland and Patzkowsky, 1996, 2008). Abrev. AF, Appalachian structural front; TG, Timiskaming Graben; OBG, Ottawa-Bonnechère graben. (B) Geological elements of the Ottawa Embayment (OE). The Ottawa-Bonnechère graben (OBG) fault distribution, with the Gloucester (GF) and Rideau (RF) faults highlighted, illustrates a west-directed curvature (from E-W to NW-SE). The OE is delimited from the adjacent western Quebec Basin (wQB) by the Beauharnois-Oka anticline, and from the northern Appalachian Basin (nAB) by the Frontenac arch (FtA). Precambran geological components include the Grenville mafic dyke swarm (green), and the NW-SE-oriented distribution of defined Proterozoic (Grenvillian) lithotectonic terrains. Definitions of possible extension of the MAH-FA boundary (dashed outline) beneath the OE stratal succession, and possible isolated block of the MAH terrain, are based on aeromagnetic geophysical patterns (Gupta 1991). Distribution of Lower Paleozoic megasequences subdivides the foreland Tippecanoe I Megasequence into an initial platform-interior, then basin-fill succession. Map is modified from Oruche et al. (2019).
embayment coeval with long-term eustasy and climate change through the Middle to Late Ordovician. Correlation with the Appalachian Basin allows for differentiation of eustatic and tectonic controls.

**Structural and Lithostratigraphic Frameworks**

The orientation of the Ottawa Embayment positioned along the southern margin of the Canadian Shield, which is underlain by Precambrian rocks, is a product of erosion and structure. The embayment is delimited from the western Quebec Basin by the Oka-Beauharnois arch, and its northern and southwestern margins are defined by uplifted basement rock underlying the Laurentian and Frontenac arches, respectively (Fig. 4.1B; Sanford, 1993). The embayment’s axis coincides with the axial trace of the intracratonic Ottawa-Bonnechère graben (Fig. 4.1; Kay, 1942), a long-lived feature beginning as an aborted Neoproterozoic continental rift (Burke and Dewey, 1972; McCausland et al., 2005) followed by Paleozoic and Mesozoic faulting in response to distal orogenesis, Mesozoic heating and uplift, and Quaternary neotectonics (Burke and Dewey, 1973; Crough, 1981; Dix and Molgat, 1998; Dix et al., 1998; Rimando and Benn, 2005; Dix and Al Rodhan, 2006; McCausland et al., 2007; Ma and Eaton, 2007; Dix and Al Dulami, 2010; Dix and Jolicoeur, 2011; Roden-Tice et al., 2005, 2012; Hardie et al., 2017; Lowe et al., 2017, 2018).

The embayment is underlain by Cambrian through Upper Ordovician strata that constitute local expressions of the cratonic Sauk and Tippecanoe I megasequences (Fig. 4.1B; Sloss, 1963; Fritz et al., 2012). The former characterizes platform-interior deposits of the southern Laurentian trailing margin (Lavoie et al., 2012; Lowe et al., 2017).
Subsequent foreland (Tippecanoe I) deposition is separated into a Middle (Whiterockian) to Upper Ordovician (Katian, Ka1) platform succession (Fig. 4.2) followed, after platform foundering (Lavoie, 1994), by a basin-fill succession (Billings Formation; Fig. 4.2) that is part of eventual siliciclastic overfilling in response to northward migration of a regional orogen-derived clastic wedge (Sanford, 1993). The youngest preserved strata in the embayment form coastal red-bed facies (Queenston Formation) of Richmondian (mid-Cincinnatian or Ka3-Ka4) age (Fig. 4.2).

**Methodology**

**Data Sources**

Sequence-based facies successions (Table 4.1) and interpreted changes in base level (Fig. 4.2) are compiled from published and unpublished (thesis) sources (see below) along with observations derived from our study. Geographic coordinates of localities are provided in Appendix A along with equivalent locality notations used in prior principal publications. Selected stable carbon and oxygen isotope profiles are used to resolve local diagenetic attributes across selected formation contacts. These data, along with information about analytical methods, standard error, and standards are reported in Oruche et al. (2019), or Chapter 3 of this thesis.

**Facies and Sequence Stratigraphy Models**

The foreland-platform succession of the Ottawa Embayment records changing abundances of coastal siliciclastics, evaporites, environmentally restricted to normal-
**Fig. 4.2A:** Stratigraphy of the Middle through Upper Ordovician foreland platform and basin succession of the Ottawa Embayment illustrating sequence stratigraphic divisions, paleoenvironmental change, and chronology of distal orogen-based tectonic events. The Millbrig Bentonite (M) and other prominent bentonites are indicated (red bar). Lithic patterns and symbols are explained in the accompanying Fig. 2B. Chronostratigraphic and biostratigraphic nomenclatures are from Webby et al. (2004) and Ogg et al. (2016), and the age of the Millbrig Bentonite is from Oruche et al. (2018). Abbreviations: Tu, Turinian; Ch, Chatfieldian; Ed, Edenian; Ma, Maysvillian; Ri, Richmondian; L, land; I, inner ramp; M, middle ramp; O, outer ramp; B, basin. The term megasequence follows Fritz et al. (2012). The double arrow for the Sauk-Tippecanoe I boundary reflects placement (a) based on interpretations prior to 2006, and (b) following Dix and Al Rodhan (2006) and this study. Formation boundaries indicated by a dashed line indicate uncertainty of age. Explanation of patterns and symbols is provided in Fig. 2B.
<table>
<thead>
<tr>
<th>Era</th>
<th>Period</th>
<th>Stage</th>
<th>Substage</th>
<th>Time (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Middle</td>
<td>Ordovician</td>
<td>Dapingian</td>
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<tr>
<td></td>
<td></td>
<td>Darrilwan</td>
<td></td>
<td></td>
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<tr>
<td>Upper</td>
<td>Ordovician</td>
<td>Whiterockian</td>
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<td></td>
<td></td>
<td>Mohawkian</td>
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<td></td>
<td></td>
<td>Cincinnati</td>
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<td></td>
<td></td>
<td>&quot;Chazyen&quot;</td>
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</table>

**Sauk Platform**

- Salt Creek Formation
- H. salina Formation
- H. hebelodontes Formation
- E. pachyura Formation
- C. swed Formation
- P. evelae Formation
- P. compressa Formation
- R. thoracica Formation
- C. georgianus Formation
- A. compressa Formation
- A. penicillata Formation
- A. penicillata Formation
- A. compressa Formation

**Shale Basin**

- Shale of the Sauk Platform

**Depositional Environments**

- TST (Transgressive Systems Tract)
- RST (Regressive Systems Tract)

**T.R. sequences**

- II
- III
- IV
- V
- VI
- VII
- VIII
- IX

**Sedimentology**

- Lithostratigraphy
- Biostratigraphy
- Depositional History
**Fig. 4.2B**: Legend for lithic patterns, symbols of fossils and sedimentary structures, and grain size, and sequence stratigraphic components relevant to Fig. 4.2A and subsequent figures.
Table 4.1: Depositional sequences, their constituent formations, sedimentary characteristics, and interpreted paleoenvironments.
<table>
<thead>
<tr>
<th>Sequence</th>
<th>Formation</th>
<th>Boundary types</th>
<th>Lithofacies Succession</th>
<th>Fossils/Grains</th>
<th>Sedimentary Structures</th>
<th>Paleoenvironments</th>
<th>Sources</th>
</tr>
</thead>
<tbody>
<tr>
<td>IX</td>
<td>Queenston</td>
<td>top: 3b</td>
<td>Interbedded red/grn-gry</td>
<td>c, br, s, o: within</td>
<td>laminae/thin beds;</td>
<td>peritidal; basal</td>
<td>a, f, g</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>sh and fgr-ss</td>
<td>basal carbonate</td>
<td>bioturbation</td>
<td>skeletal-oooid shoals</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Carlsbad</td>
<td>top: 1</td>
<td>gry to grn sh, fgr-ss, s-o pkst, s- grnst; thin bent beds</td>
<td>c, b, br, g, bry, na, gr</td>
<td>graded bedding; parallel- and cross-lamination; burrows and burrow mottling</td>
<td>bathymetric gradient; shallowing from storm wave base</td>
<td>g</td>
</tr>
<tr>
<td></td>
<td>Billings</td>
<td>top: 1</td>
<td><strong>upper division:</strong> drk gry clysh with sltst and fgr-ss laminae to thin beds</td>
<td>gr, br, t, ce, na</td>
<td>parallel- and cross- lamination, gutter casts</td>
<td>anoxic to dysoxic disrupted by turbidity flows</td>
<td>e, g, h</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td><strong>bentonite</strong></td>
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<td></td>
<td></td>
<td></td>
<td><strong>lower division:</strong> blk clysh, rare sltst laminae</td>
<td>gr, b, na: pyritic</td>
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<tr>
<td>VIII</td>
<td>Lindsay</td>
<td>top: 1</td>
<td><strong>Eastview Mbr:</strong> nodular lmd with sh grading up into interbedded lmd and sh with lenses/beds of skel rdst</td>
<td>c, b, br, bu, t c, b, br, t,</td>
<td>laminated sh and rdst; massive lmd</td>
<td>fluctuating oxic to dysoxic; storm and hiatal skeletal concentrations</td>
<td>a, j, n</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td><strong>Nepean Point Mbr:</strong> nodular s pkst, wkt, grnst; with sh</td>
<td>c, g, b, br, t, na, rc, ac-d</td>
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<tr>
<td>VII</td>
<td>Verulam</td>
<td>top: 1</td>
<td>interbedded lmd, s-wkt and s-grnst, clc-sh; bentonite</td>
<td>br, c, b, g, bry, o, t, bu, bi</td>
<td>graded beds; HCS; scour features</td>
<td>mid- to outer ramp, deepens, then shallows</td>
<td>a, i, j</td>
</tr>
<tr>
<td>VI</td>
<td>RST</td>
<td>Hull</td>
<td>H3</td>
<td>top: 2</td>
<td>a) predominantly: s-grnst/rdst and s-sh</td>
<td>c, co, cT, p, sp, n, m, bry</td>
<td>thick beds; crossbeds; reverse flow; amalgamated bedding, erosional surfaces</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Rockland</td>
<td>H2</td>
<td>top: 2</td>
<td>lozenge-shaped fgr c-grnst, with sh laminae and partings</td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>H1</td>
<td>top: 2</td>
<td>thin interbeds of s-rdst / grnst and s-rich sh</td>
<td>ca-Solenopora, b, br, bry c, n, t</td>
<td>Large scale crossbeds and primary dip (SSW: 10–20°)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>base: 3</td>
<td>rhythmic interbeds of lmd and clc-sh, and less common s grnst/pkst</td>
<td>br, bry, c, g, cr, n, o, ac-c</td>
<td>gradational lithofacies contacts</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>base: 5</td>
<td></td>
<td></td>
<td>platform division: low-energy normal marine bank</td>
</tr>
<tr>
<td>TST</td>
<td>L’Orignal</td>
<td>top: 3</td>
<td>upsection change: upper: rhythmic interbeds of s pkst (or lmd) and clc-sh</td>
<td>c, bu</td>
<td></td>
<td>normal-marine transgression from shallow shoreface to subtidal inner ramp</td>
<td>b</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>lower to mid: s-pkst/grnst</td>
<td>t, a, acd, b, br, c, bry, g, o, m, sp,</td>
<td>load structures</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>base: 3</td>
<td>base: qa, with rare blackened phosphatic lithoclasts (qa, wke)</td>
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</tr>
<tr>
<td>V</td>
<td>RST</td>
<td>Watertown</td>
<td>top: 4</td>
<td>W2: thick beds; s- and p-wkst; locally nodular fabric</td>
<td>b, br, c, co, g, n, t br, bry, c, co, g, n, t, s</td>
<td>pervasive 3-D burrowing; hardgrounds; borings skeletal fragmentation</td>
<td>agitated, subtidal normal-marine ravinement deposit</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>base: 3</td>
<td>W1: cgr s-pkst, -grnst, -rdst</td>
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<td></td>
<td>RST</td>
<td>TST</td>
<td>top: 3</td>
<td>L2: lmd (minor s-pkst, -rdst, -flst) Tetradium mud-rich to poor mounds and capping biostrome</td>
<td>b, bry, cT, g</td>
<td>differential compaction; storm deposits</td>
<td>protected inner ramp</td>
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<td>IV</td>
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<td></td>
<td></td>
<td>L1: s/o-pkst/grnst; s-lmd; clc-sh</td>
<td>b, br, bry, c, co, g, n, t</td>
<td></td>
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<tr>
<td></td>
<td>Lowville</td>
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<tr>
<th></th>
<th>RST</th>
<th>TST</th>
<th>top: 3</th>
<th><strong>Division IV</strong>: fgr-qa; dmd; lmd; e; sh</th>
<th>bry</th>
<th>dessication cracks; planar bedding; evaporite nodules; burrows</th>
<th>arid, inter-/supratidal</th>
<th>a, c</th>
</tr>
</thead>
<tbody>
<tr>
<td>III</td>
<td></td>
<td></td>
<td></td>
<td><strong>Division III</strong>: lmd; lt-pkst-rdst; s-mdst/wkst; sh</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td><strong>Division II</strong>: small/large m-mounds, s-pkst/grnst; clc-sh</td>
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<td></td>
<td><strong>Division I</strong>: clc-sh; sh; fgr-ss (phosphatic); lmd; s-o pkst/grnst</td>
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<tr>
<td></td>
<td>Pamela</td>
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<tr>
<th></th>
<th>RST</th>
<th>TST</th>
<th>top: 2</th>
<th>thick beds of d/lmd, and thin clc-sh; local thin beds of s-rdst clc-sh with rare thin beds of lmd, vfgr-ss, ps-rdst, s-cg</th>
<th>br, b, o, t, c (rare)</th>
<th>skeletal beds with packed fabric</th>
<th>low-energy platform: storm deposits or productivity pulses</th>
<th>a, k</th>
</tr>
</thead>
<tbody>
<tr>
<td>II</td>
<td>Hog's Back</td>
<td></td>
<td></td>
<td></td>
<td>brl</td>
<td>load cast; brl-rich (phosphatic) layers; disrupted bedding; ripple cross-lamination</td>
<td>inner platform shale basin; shallowing/deepening events</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>basal vfgr-ss capped by burrowed omission surface</td>
<td></td>
<td>burrows filled with brl-grnst</td>
<td>transgressive deposit</td>
<td></td>
</tr>
<tr>
<td>I</td>
<td>RST</td>
<td>Rockcliffe</td>
<td>top: 3</td>
<td>interbeds: fgr-qa and sh; sh-u/sa-u; qa-lenses; brl-rich beds; channel-fill cgr-qa near base, with brl debris</td>
<td>brl; bu (seaward gradient from Cruziana to Skolithos ichnofacies)</td>
<td>planar bedding, shale rip-up clasts; cross-bedding, reverse flow; truncation surfaces; synsedimentary deformation;</td>
<td>tidal- and wave-dominated estuarine setting</td>
<td>a, d, l, m</td>
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<tr>
<td>TST</td>
<td>Providence Island</td>
<td>top: 1-3</td>
<td>thin interbeds: d/lmd, clc-sh, qa; s-pkst, lt-dmd rdst, rare s-grnst; local e-beds/nodules; mm-mounds; medial qa marker bed; more shale above marker bed lowermost few metres: dmd fragments similar to Fort Cassin Fm; m-mounds; qtz pebbles</td>
<td>rare br, g</td>
<td>synsedimentary deformation, fractures; dessication cracks; local burrows; cross-lamination in qa, reverse flow indicators</td>
<td>low-energy but wave- and tide- influenced arid peritidal platform</td>
<td>a, d</td>
<td></td>
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</tbody>
</table>

**Boundary type:** 1, conformable; 2, abrupt (does not include evidence of erosion or stylolite); 3, erosional: a, local; b, regional; 4, stylolitic

**Lithology:** sh, shale; clysh, clayshale; ss, sandstone; md, mudstone; lmd, lime mudstone; dmd, dolomudstone; wkst, wackstone; pkst, packstone; grnst, grainstone; rdst, rudstone; qa, quartz arenite; wke, wacke; e, evaporite (bed, nodules); qtz quartzite.

**Modifiers:** Grain Size: fgr, fine-grained; mgr, medium-grained; cgr, coarse-grained. Texture/composition: clc, calcareous; lt, lithoclastic; sh-u shalying upwards; sa-u sandying upwards.

**Colour:** gry: grey, grn: green; blk: black

**Sedimentary structures:** HCS, hummocky cross-stratification

**Fossils/Grains:** a, calcareous algae (ca: calcareous; n: non-calcareous; -d, dasyclad; -c, codiacean); b, bivalve; bo, borings; br, brachiopod; brl, lingulid brachiopod; bry, bryozoan; c, crinoid; c: coral (cT, Tetradium: co, other colonial forms; cr, rugose); g, gastropod; gr, graptolite; m, microbial (type defined); n, nautaloid; o, ostracodes; oo, ooids; p, pellets/peloids; s, skeletal; s-o, skeletal-ooid; st, stromatoporoids; sp, sponge/spine spicules; tr, trilobite.

**Sources:** a, this study; b, Oruche et al., 2018; c, Salad Hersi and Dix 1999; d, Dix and Al Rohan, 2006; e, Dix and Joliceour, 2011; f, Sharma and Dix, 2004; g, Sharma et al., 2003; h, Sharma et al. 2005; i, Kiernan, 1999; j, Gbadeyan and Dix, 2013; k, Salad Hersi and Dix 1997; l, Hofmann 1979; m, Troyer, 1979; n, Williams, 1991
marine carbonates, and near-shore to outer platform shale (Table 4.1). Such variation fits a lateral variation expected along a tropical epicontinental carbonate ramp (Fig. 4.3A; Handford and Loucks, 1993), but one marginal to an expanding deep-ocean basin in response to orogenic development (Lavoie, 1994). There is no evidence for a distally steepened margin within the basin, and coeval ramp geometries are well represented elsewhere along the contemporary southern Laurentian foreland (Pope and Read, 1998; Brett and Baird, 2002; Lavoie, 2008). During the Middle through Late Ordovician, the Ottawa Embayment lay within southern lower latitudes (Fig. 4.1A; Torsvik and Cocks, 2017). A second facies model, one that describes siliciclastic tide- and storm-influenced estuarine or fan-delta settings (Fig. 4.3B; Dalrymple and Choi, 2007; Shchepetkina et al. 2018), serves to characterize a period of extensive siliciclastic cover (Rockcliffe Formation) restricted to the Middle Ordovician (Fig. 4.2).

In sequence stratigraphic analysis, the depositional sequence (Mitchum et al., 1977) constitutes the primary mapping unit defined by spatial and temporal patterns of facies successions relative to paleosurfaces of exposure, conformity, or regression/transgression (Catuneanu, 2006). The above depositional models help differentiate changes in local water depth through time (Fig. 4.2). Water depth is a function of sediment supply or autogenic (carbonate) production along with effects of subsidence or uplift, and global sea level variation related to climatic and-or lithosphere dynamics including ocean-basin spreading rates, deformation, and mid-ocean ridge volumes (Kearey and Vine, 1992). Of available models (see Catuneanu, 2006), the transgressive-regressive (T-R) sequence model (Fig. 4.3C) of Embry (1993, 1995, 2009) proves to be the most practical for the
**Fig. 4.3:** Generalized models used in sequence stratigraphic analysis.  
(B) Salinity variation and current directions (arrows) along an estuarine setting (inset) with specific sedimentary and ichnofacies associations illustrated in a cross-sectional profile for the Middle Ordovician Rockcliffe Formation (Ottawa Embayment) grading seaward into increased limestone content (blue) of the Laval Formation in the western Quebec Basin, forming platform-interior equivalent successions of the lower Chazy reef succession outcropping in the Vermont region. Based on Hoffman (1979), Dalrymple and Choi (2007), and Dix et al. (2013).  
(C) Depositional elements of the Transgressive-Regressive (T-R) sequence stratigraphic model. Modified from Embry (2009). Abbreviation: TST, Transgressive system tract; RST, Regressive system tract; MRS, Maximum regressive surface; SU, Subaerial unconformity; SR, Shoreline ravinement.
Ottawa Embayment. A T-R sequence is bounded by subaerial unconformities along the basin margin but maximum regressive surfaces towards the basin center (Fig. 4.3C). Absence of platform-margin sections in the embayment precludes characterizing sea-level position (highstand, lowstand) relative to the continental margin. Instead, shallowing (S)- or deepening (D)-upward successions (Fig. 4.2) signal shifts in the shoreline (Embry, 2009).

**Sequence Stratigraphy: Ottawa Embayment**

Interpreted transgressive-regressive cycles, nature of stratigraphic boundaries, and paleoenvironmental succession are summarized in Fig. 4.2 and Table 4.1. Here, sequences are briefly described followed by interpretation of paleoenvironmental conditions.

**Sauk-Tippecanoe I Megasequence Boundary**

The Sauk-Tippecanoe I megasequence boundary was first placed at the prominent lithic boundary separating dolostone of the Lower to Middle Ordovician Beekmantown Group from overlying sandstone of the Middle Ordovician Rockcliffe Formation (a in Fig. 4.2; Sloss, 1963; Sanford, 1993). The contact was subsequently interpreted to document a hiatus related to a eustatic lowstand (Barnes et al., 1981; Barnes, 1984). Dix and Al Rohan (2006) proposed that the boundary should be repositioned within the upper Beekmantown Group (b in Fig. 4.2) at the now defined Fort Cassin-Providence Island boundary (Landing and Westrop, 2006), which marks a change in both regional tectonic and oceanographic frameworks across the embayment and western Quebec Basin (Bernstein, 1991). In the Ottawa Embayment, this change is characterized by regional
**Fig. 4.4:** Stratigraphic relationships of the Providence Island Formation, Ottawa Embayment. (A) A generalized cross section of the formation along an axial transect within the Ottawa embayment (modified from Dix and Al Rodhan, 2006). The section illustrates a west-directed onlap across faulted platform of the Fort Cassin Formation. Additional sedimentary attributes are indicated. D-S patterns are defined in Fig. 4.2B. (B) Field photograph of the Fort Cassin (a) and Providence Island (b) formations at Loc. 25. Fractures (arrows) are filled with brown dolostone of the Providence Island Formation. Pen for scale. (C) Thin-section photograph (plane-light) of the gastropod rudstone that caps the Fort Cassin Formation at Loc. 25. Vadose-like textures of dolomicrite (arrow) are associated with dolomitized casts of gastropods (a) and overlain by later dolomite cement (b). Scale bar is 1 mm. (D) Thin-section photograph (plane-light) of erosional boundary (white dash line) at Loc. 25 that superposes Providence Island strata (b) on truncated diagenetic fabric within the Fort Cassin (a) Formation. Scale bar 1 mm.
west-directed onlap of the Providence Island succession across a faulted Fort Cassin paleoplatform (Fig. 4.4A), by pervasive syndepositional deformation within the Providence Island succession (Fig. 4.4A; Dix and Al Rodhan, 2006), and, by an abrupt transition to shallower and oceanographically more restricted conditions (Bernstein, 1992).

Two closely spaced (~50 m) localities (Loc. 24 and 25: see inset map in Fig. 4.4) separated by a present-day fault (Dix et al., 1998) illustrate contrasting sedimentary attributes of this transition. At Loc. 25, the upper two metres of the Fort Cassin Formation record a gradational upsection change from bioturbated crinoidal mesocrystalline dolostone into gastropod-lithoclast dolorudstone. Downward-tapering (< 30 cm) synsedimentary fractures that extend into the underlying bioturbated dolostone (Fig. 4.4B) are filled by dolostone contiguous with overlying Providence Island strata. Gastropod forms (Fig. 4.4C), some with preserved shells, others occurring in the form of steinkerns, molds, and lithoclasts, are typically < 5 mm in size but are of obscure types (see Appendix I) due to incomplete preservation. Skeletal remains as well as Inter- and intraskeletal matrices are dolomitized, and at least two stages of isopachous dolomite cements post-date a meniscus-like fabric of intergranular dolomicrite (Fig. 4.4C). All the above dolomitization predates Providence Island deposition that overlies the erosionally truncated diagenetic fabric (Fig. 4.4D).

In contrast, Loc. 24 occurs within ~50 m of the first locality. The equivalent interformational contact truncates burrowed micro- to mesocrystalline crinoidal dolostone similar to that low in the section at Loc. 25. Lithic fragments in overlying Providence Island strata are derived from the underlying Fort Cassin strata (Fig. 4.4A), but quartzite
pebbles to cobbles also occur along diastemic erosional boundaries. Two three-dimensional burrow fabrics have been truncated by the formational disconformity (Fig. 4.5A, B; see Appendix I for detail). Burrow margins are typically darkened (Fig. 4.5C), but not pyritic. Burrow fill is not Providence Island sediment.

Stable isotope ($\delta^{13}C$, $\delta^{18}O$) profiles (Fig. 4.5D) reveal a prominent negative shift in both $\delta^{13}C$ and $\delta^{18}O$ within the gastropod-lithic rudstone at Loc. 25. Such a compositional shift mimics that of sub-disconformity meteoric alteration in Quaternary carbonates with incorporation of atmospheric and/or soil-derived carbon (Allan and Matthews, 1982), and is similar to other examples of meteoric alteration interpreted from Ordovician platform successions (Tobin et al., 1999). At Loc. 24a, a coupled negative isotopic shift is absent within the Fort Cassin burrowed facies, and there is a pronounced difference in the magnitude of both isotopic ratios with $\delta^{13}C$ and $\delta^{18}O$ values more negative (~2 ‰) in the upper Fort Cassin strata at Loc. 24a compared to Loc. 25. The amount of differential erosion between these two sites is at least the thickness of the gastropod rudstone at Loc. 25, or ~10s of centimetres.

The above data suggests there occurred post-Sauk exposure, fracturing, and development of meteoric vadose and phreatic diagenesis at Loc. 25 (Flügel, 2010; Tobin et al., 1999). In close proximity (~50 m), such evidence is lacking for Loc. 24. However, possible ichnofossil tiering illustrates a fabric transition from initial thorough excavation of the subsurface to discrete small burrows (Fig. 4.5A, B). This change may reflect increasing restriction in oceanographic conditions, especially oxygen (Bromley, 1990). These two
**Fig. 4.5:** Petrographic and lithic characteristics of the Fort Cassin-Providence Island formation boundary. (A) Photograph of polished rock slab (left: uninterpreted; right: interpreted) showing the disconformity (highlighted in black) that truncates burrowed dolostone (arrow highlighted in black) of the Fort Cassin Formation at Loc. 24a. Abbreviation: H, host rock; B, burrow-fills; S, skeletal fragment; dashed line highlights a possible U-shaped burrow; horizontal burrows highlighted with white arrow; a, lithoclasts of Fort Cassin lithofacies. Scale bar = 1 cm. (B) Photograph of a polished rock slab illustrating the Fort Cassin-Providence Island disconformity (highlighted in black) at Loc. 24a, alignment of apparent horizontal burrows along a slope (defined by dashed line) within Fort Cassin dolostone. (C) Photograph of polished rock slab showing darkened boundary of horizontal burrow in Fort Cassin strata at Loc. 24. Scale bar = 1 mm. (D) Lithostratigraphy and chemostratigraphic profiles ($\delta^{13}$C, $\delta^{18}$O) across the Fort Cassin-Providence Island formation contact at localities 24 and 25. Explanation of D-S patterns and related symbol is provided in Fig. 4.2B.
sites suggest development of an irregular post-Sauk paleobathymetry over distances of
tens of metres, possibly local islands and shallowing marine lagoons. Such a
paleogeography may indicate that the present-day fault separating these localities reflects
an inherited syndepositional structure that controlled post-Sauk bathymetry.

**Sequence I (Providence Island-Rockcliffe formations)**

*Transgressive system tract*

West-directed onlap of the Providence Island succession extends across the step-like
faulted basement of the Fort Cassin succession (Dix and Al Rodhan, 2006) and onto
Precambrian metamorphic rocks northwest of the embayment (Fig. 4.4A; Wilson, 1924;
Bernstein 1992). The formation consists of thinly interbedded siliciclastics, local
evaporites, and fine-grained carbonates (mostly dolostone) that identify a dynamic, yet
arid peritidal platform influenced by waves and tides (Table 4.1; Dix and Al Rodhan,
2006). We are uncertain about details of vertical change in water depth in the lowermost
formation except immediately beneath a regional marker bed of sandstone (Fig. 4.4A)
that illustrates seaward progradation of shore-derived siliciclastics (Dix and Al Rodhan,
2006). Above this unit, there is a more shale-rich succession interpreted to document
deeper, but still shallow subtidal conditions extending into the western Quebec Basin
(Bernstein, 1991; Hoffman and Bolton, 1998). We interpret the formation to document
net deepening.

*Regressive systems tract*

A subsurface transect of closely-spaced borehole gamma-ray profiles (Fig. 4.6A; Dix et
al., 2013) from the central embayment suggests that erosional contacts defining the
Fig. 4.6: Field-scale geophysical and lithic attributes of the Rockcliffe Formation (A) Stratigraphic correlation along transect A-A’ using gamma-ray well logs of the uppermost Providence Island Formation (orange-brown) and sub-units within the Rockcliffe formation: olive green, channel-fill facies; green, interbedded siltstone and sandstone; yellow, more sand rich to sand dominated intervals. Right-directed arrows define shalying upwards; left directed arrows indicate sanding upwards. Faults are interpreted to accommodate offset gamma ray signatures and tie-lines. (B) Field photo and lithostratigraphic sections showing the erosional to abrupt boundaries between the Providence Island (a) and Rockcliffe (b) formations at Loc. 28 and 26. (C) Photograph of polished rock slab illustrating fine-grained (b) lithofacies of the Rockcliffe (R) Formation draped down into a fracture extending into the Providence Island (Pl) Formation at Loc. 37. Coarse-grained channel-fill-like facies (b) is partially disrupted by the downward draping. Scale bar = 1 cm. (D) Photograph of polished rock slab illustrating the basal channel-fill facies of the Rockcliffe Formation (R) disconformably overlying Providence Island strata. There is abundant brachiopod (lingulid) shell debris, a large lithoclast (white arrow) of the reworked basal fine-grained lithofacies, and an erosional boundary (black arrow) within the channel-fill facies on which cross-bedded strata developed. Scale bar = 2 cm. (E) Lithostratigraphic section at Loc. 5 (see Fig. 4.6A) of the Rockcliffe Formation showing repetition of shaling- (sh-u) and sanding (s-u) successions at Loc. 5. Section is measured in metres.
Providence Island-Rockcliffe formation boundary (Fig. 4.6B), emphasized previously as evidence for significant hiatus, are part of structurally controlled channelization of km-scale width that cut down no more than a few metres into the Providence Island succession. The channel-fill succession consists of coarse-grained, locally cross-bedded, quartz and feldspathic sandstone to granule conglomerate with abundant phosphatic shell fragments of lingulid brachiopods (Wilson, 1946; Dix and Molgat, 1998). Beyond the channel margins, however, interbedded to laminated fine-grained sandstone and coarse siltstone forms the basal lithology of Rockcliffe Formation that is locally deformed by synsedimentary fractures extending down into Providence Island dolomudstone (Fig. 4.6D; Dix and Molgat 1998). Overlying this basal succession and involved in local deformation is coarse-grained channel-like sandstone with lingulid shell debris (Fig. 4.6D). There are also pebble-size lithoclasts in the coarse-grained sandstone (Fig. 4.6D) similar to the fine-grained basal facies succession beyond the channel margins. These features suggest that channel formation reworked an existing basal Rockcliffe sediment. Thus, local fracturing may be contemporaneous with channelization, and both post-date onset of Rockcliffe accumulation. The characteristic erosional contact so prominent in outcrop (Fig. 4.6B), therefore, reflects structural control.

There is little published sedimentological analysis of the Rockcliffe Formation. The unit forms a siliciclastic platform-interior to the more seaward lower Chazyan reefal succession (Fig. 4.3B) in the Champlain Valley of New York State and Vermont (Hofmann, 1979; Dix et al., 2013). Hofmann (1979) interpreted the Rockcliffe Formation as being part of a coastal to shallow-water tide- and/or wave-dominated estuarine (brackish to marine) setting, with trace-fossil assemblages characterizing a seaward
gradient, from *Skolithos* to *Cruziana* ichnofacies (Fig. 4.3B), toward the central embayment. Other studies report metre-scale-thick sand bodies, successions of interlaminated sandstone and siltstone, shale abruptly overlying sandstone, and low-angle truncation surfaces that may be of structural or storm influence (Wilson, 1946; Hofmann, 1979; Dix et al., 2013). A core section (Loc. 5: see inset map of Fig. 4.6) exhibits stacked, ~44 metres, muddying-upwards successions capped by a sandying-upward interval (Fig. 4.6E; Table 4.1). Lingulid brachiopods are locally common to abundant throughout the formation. Gamma-ray well logs (section A-A’, Fig. 4.6) contain bell- and funnel-shaped half-profiles associated with sand-dominated intervals (Fig. 4.6A). Such patterns are commonly interpreted to define channel-fill and progradational shoal systems, respectively (Cant, 1992). Based on Hofmann’s (1979) interpretation, therefore, we interpret the Rockcliffe succession to represent deposition within an outer estuary or broad fan-delta setting (Fig. 4.3B; Dalrymple and Choi, 2007) whereas structurally controlled channelization may document a short-lived seaward jump of the fluvial-tidal transition (Shchepetkina et al., 2019).

Transformation from the peritidal Providence Island succession to brackish facies of the Rockcliffe Formation follows increased shale content in the former formation, and replacement of a dolostone-siliciclastic tidal microfabric by wholly siliciclastic tidal microfabric (Dix and Al Rodhan, 2006; Dix et al., 2013). The maximum flooding surface occurs with appearance of the initial fine-grained Rockcliffe facies otherwise reworked by structurally-controlled channelization (Fig. 4.4). Excluding this latter event as a structural artifact, we cannot distinguish sandying- and muddying-upwards successions (Fig. 4.6A) as allogenic versus autogenic (lateral migration) origins based on available
data. This translates into considerable uncertainty about deepening versus shallowing trends within the formation, as we illustrate in Fig. 4.2.

**Sequence II (Hog’s Back Formation)**

The Hog’s Back Formation (Salad Hersi and Dix, 1997) thickens eastward through the embayment (Salad Hersi and Dix, 1997) and grades into the middle to upper part of the Chazyan reefal succession to the east (Dix et al. 2013). At Loc. 27, the formation consists of a lower heterolithic shale-dominated division succeeded by a thick-bedded carbonate-rich upper division (Fig. 4.7A; Salad Hersi and Dix, 1997). Within 10 kms of this site (Loc. 38: see inset map of Fig. 4.7) the lower division is absent, and the upper carbonate-rich succession onlaps a disconformity-bound wedge of synsedimentary disrupted Rockcliffe strata (Dix et al., 2013). This demonstrates regional deposition of the Hog’s Back Formation across a paleosurface marked by local synsedimentary structural deformation.

*Transgressive systems tract*

At Loc. 27, a thin bed of fine-grained quartz arenite disconformably overlies the Rockcliffe Formation and is abruptly overlain by the remainder of the lower shale succession (Fig. 4.7B; Salad Hersi and Dix, 1997; this study). The top of the arenite bed exhibits abundant horizontally branched and vertical thallasinoid-like burrows packed with shell fragments of lingulid brachiopods (Fig. 4.7C, D). The zone of bioturbation disappears downsection over 5-10 cm (Fig. 4.7D). The arenite is interpreted as a transgressive sand (Catuneanu, 2006) resulting from transgressive reworking of unlithified Rockcliffe sediment during transgression. Its burrowed top defines a
Fig. 4.7: Field characteristics of the Hog’s Back Formation at Loc. 27. (A) Lithostratigraphic section of the Hog’s Back Formation, Loc. 27, with D-S pattern and symbols defined in Fig. 4.2B. (B) Outcrop photograph showing the disconformity (black line) separating Rockcliffe (a) strata from the basal transgressive sand (b) of the Hog’s Back Formation; the lower shale succession is exposed in the background. Hammer for scale. (C) Photographs of polished rock slabs (Loc. 27) illustrating cross-sectional form (upper) of burrows extending down from the top of the basal sandstone bed, and (lower) bedding plane geometry. The burrows are packed with shell fragments of lingulid brachiopods. Scale bar = 1 cm. (D) Photograph of bedding plane exposures of successive beds down through the transgressive sandstone illustrating abrupt reduction in bioturbation with depth. Measuring tape (inches, centimetres) for scale. (E) Outcrop photograph showing soft-sediment deformation along a skeletal-rich limestone bed within tens of centimetres beneath the upper boundary of the Hog’s Back Formation, Loc. 27.
firmground, a bioturbated omission surface (Pearson et al., 2012) or hiatal burrow-lag concentration (Kidwell, 1993). The top of the TST may be the top of this basal bed or, as illustrated in Fig. 4.4, occurs at the highest of several recurring phosphatic shell-rich intervals that also form lag concentrations within the lower shale-rich interval (Fig. 4.7A). Alternating limestone, shale, and local siliciclastic beds may form subordinate TST-RST cycles comprising the lower shale division.

Regressive System Tract

Net shallowing transformed the shale basin into a restricted low-energy (muddy) carbonate platform with interbedded shale (Salad Hersi and Dix, 1997; Dix et al., 2013). Thin interbeds of densely packed skeletal (bivalve, brachiopod, ostracode) rudstone represent either episodic storm deposits (Salad Hersi and Dix, 1997; Desjardins, 2012) or spikes in biotic productivity related to short-term change in environmental conditions. Sedimentary boudinage and contorted lenses of skeletal-rich limestone immediately below the formation’s upper boundary (Fig. 4.7E) signal continued impact of synsedimentary deformation.

Sequence III (Pamelia Formation, Ottawa Group)

The Pamelia Formation is the lowest formal division of the Ottawa Group (Uyeno, 1974), which unites eight formations by a common, mostly limestone-shale lithic association (Fig. 4.2; Wilson, 1946). The formation thickens into the center of the embayment from both the west and eastern limits (Salad Hersi and Dix, 1999). Salad Hersi and Dix (1999) established a bipartite lithic division whereas we identify four stratigraphic divisions (Table 4.1, Fig. 4.2).
**Transgressive Systems Tract**

Divisions I and II (Table 4.1) constitute net transgression from a shale-rich inner platform to normal-marine subtidal carbonate platform. Division I is a heterolithic succession (Fig. 4.8A; Table 4.1), one that exhibits regional heterogeneity in lithic abundance and distribution. Overall, a basal shale gives way upsection to greater abundance of limestone with a restricted skeletal assemblage, and there are at least three very thick (up to metre-scale) bioturbated phosphatic arenite beds (Fig. 4.8A) that punctuate this stratigraphic interval (Fig. 4.8A; Salad Hersi and Dix, 1999; Dix et al. 2013). Division II is a relatively thin (5-7 m) composite unit of skeletal-rich limestone and shale with two prominent levels of small to large microbial (stromatolite, thrombolite) mounds (Fig. 4.8B; Nehza and Dix, 2012). The division extends regionally (100+ km) across the western embayment, and possibly deep into the craton-interior (Kang, 2018).

**Regressive Systems Tract**

Divisions III and IV make up most of formation and characterize embayment-wide uniformity in deposition shallowing from peritidal (Division III) to intertidal-supratidal (Division IV) settings (Fig. 4.8C, Table 4.1; Salad Hersi and Dix, 1999). Combined, they record the RST (Fig. 4.2). The divisions are separated by a thin evaporite-bearing quartz arenite that rests disconformably on Division III strata. Both divisions illustrate decimetre-scale carbonate-siliciclastic rhythmicity (Salad Hersi and Dix, 1999), and Division IV contains thick beds of sandstone (Fig. 4.8D) and dedolomitized intervals. The upper formation boundary is a disconformity usually capping dolostone (Fig. 4.8C) but with local sandstone forming the uppermost unit (Fig. 4.8D). The siliciclastic unit
Fig. 4.8: Field characteristics of the Pamela Formation. (A) Outcrop exposure (Loc. 27) of Division I with basal contact exposed (white arrow) and showing thick sandstone (S) and upsection increase in limestone (ls) within a host shale succession. (B) Outcrop exposure of two levels of buildups (white arrows) forming Division II at Loc. 35. (C) Outcrop exposure of dolostone (a) of Division IV overlain disconformably (black line) by limestone (b) of the Lowville Formation at Loc. 34. (D) Outcrop exposure (Loc. 7) showing two levels of sandstone (c) in Division IV dolostone (a) overlain disconformably (thick black line) by Lowville strata (b). (E) Lithostratigraphic and chemostratigraphic ($\delta^{13}$C, $\delta^{18}$O) profiles through the upper Pamela Formation and lowermost Lowville Formation at Loc. 12 and 7. D-S pattern and related symbols are explained in Fig. 4.2B.
intermediate to divisions III and IV is best placed within the latter division, marking onset of a more lithologically diverse (carbonate, siliciclastic) interior setting (Fig. 4.3A). Two examples of δ¹³C and δ¹⁸O profiles through the upper formation (Fig. 4.8E) exhibit strong isotope covariance suggesting significant diagenetic modification (Columbié et al., 2014). However, negative excursions in both ratios occur in dolostone beneath formation-top sandstone at Loc. 7 compared to only a negative shift in δ¹³C values in dolostone at Loc. 12. This difference might reflect locality-based variation in pore-water salinity (Tobin et al., 1999).

Sequence IV (Lowville Formation)

Transgressive Systems Tract

Most of the formation consists of a lower thinly interbedded succession of low to high-energy shallow-water carbonate facies (L1, Fig. 4.9A, Table 4.1), including local ooid grainstone, with minor presence of shale. The overall high-energy facies succession is interpreted as a composite transgressive succession (Fig. 4.2) extending across a differentially eroded Pamela paleoplatform.

Regressive Systems Tract

An upper division (L2: Fig. 4.9A, Table 4.1) of the formation consists of thin beds of lime mudstone, rare skeletal rudstone, and mounds to biostromes of *Tetradium* sp. (Fig. 4.9B). The lower-upper division contact is abrupt, marking an interpreted deepening into a quiet-water lagoon setting that extended across the embayment (Oruche et al., 2018). However, subsequent shallowing characterizes the upper division as an RST as represented by an upsection increase in abundance of mounds, then local development of
Fig. 4.9: Lithostratigraphy and field photos of sequence boundaries within the Lowville through Hull formation succession. (A) A correlation between Locs. 3 and 21 illustrates formation-thickness variation and segmentation of the L’Orignal paleoplatform producing coeval development of Rockland Formation and Unit H1 strata of the Hull Formation. This division is eliminated through disconformable overstepping of Unit H3 (Hull Formation) across the Rockland Formation. Datum for the correlation is the sequence boundary (thick black line) that caps the Watertown Formation. D-S patterns and related symbols are explained in Fig. 4.2B. (B) Facies association L2 of the Lowville Formation at Loc. 9 (see Fig. 4.9) showing upsection increase in Tetradium mounds (m) to biostrome (b), along with thin skeletal rudstone (arrow) that is likely a storm deposit within the host lagoonal mud (a). Pen knife (circled) for scale. (C) The Lowville-Watertown formation contact (arrow), Loc. 4, is characteristics of the planar disconformity that truncates Tetradium mounds/biostromes (a) and is overlain by skeletal rudstone to mudstone beds of Watertown Formation (b). Chisel for scale. (D) Planar to slightly undulating disconformity defining the top of the Watertown Formation (a) at Loc. 9. The erosional surface truncates 3-dimensional burrows filled with altered volcanic-ash and crinoid debris (arrow) and overlain by the L’Orignal Formation (b). Canadian quarter (23.81 mm diameter) for scale. (E) Erosional truncation (at finger) of lime mudstone of the L’Orignal Formation (a) overlain by grainstone of Unit H1 of the Hull Formation (b), at Loc. 6.
biostromes that include local high-energy shoal facies (Wilson, 1946; Salad Hersi and Dix, 1999). The *Tetradium*-bearing biofacies is part of an interbasinal gradient extending into the northern Appalachian Basin where coeval *Tetradium*-microbial mounds occur in southern Ontario, then microbial mounds farther south in New York State (Oruche et al. 2018). In the embayment, the top of the sequence is a disconformity defined by planar truncation of mounds or biostromes (Fig. 4.9C).

**Sequence V (Watertown Formation)**

*Transgressive Systems Tract*

This formation defines marked thickening into the central embayment (Oruche et al., 2018) as illustrated by the contrast between localities N and T (Fig. 4.9A), yet both localities denote truncation of underlying Lowville strata. Apart from Loc. 21, and other sites in the eastern embayment, basal deposits (W1: Fig. 4.9A, Table 4.1) of coarse-grained, fragmental skeletal-rich limestone (Table 4.1) thicken (<2 m) to the west (Oruche et al., 2018). Loc. 21 is the only site east of the Gloucester Fault in which grainstones (similar to the W1 succession) appear higher in the formation capping an initial mud-dominated succession (Fig. 4.9A). Regionally, subdivision W1 is interpreted to form a transgressive deposit related to west-directed transgression across the horizontally-planed Lowville biostromes (Fig. 4.9B) and into an overall shallower setting west of the Gloucester Fault than to the east (Oruche et al., 2018, 2019). Loc. 21 is located within a zone of fault splays on the eastern side of the Gloucester Fault (Williams et al., 1984). The appearance of grainstones at Loc. 21 may reflect shoaling related to changing subsidence across the structural transition from east to west.
Regressive Systems Tract

Most of the Watertown Formation consists of thickly bedded (tens of centimetres to metre), well burrowed peloidal/skeletal wackestone and mudstone (W2, Fig. 4.9A; Table 4.1). The facies is of regional extent, and considered typical of this formation (Barnes 1967; Cameron, 1968; Cameron and Mangion, 1977; Oruche et al., 2018). However, this succession thins dramatically (14+ to <4 m) into the western embayment (Loc. 9, see inset map in Fig. 4.9) accompanied by increased frequency of planar to undulating hardgrounds (Oruche et al., 2018). The W2 facies succession characterizes an agitated subtidal setting (Barnes, 1967; Oruche et al., 2018), but the west-directed thinning and hardground development denotes that it was episodically punctuated by wave planation and cementation possibly allied with lowered sea level (Oruche et al., 2018). Such a rhythmic succession, nonetheless, characterizes net aggradation across the platform, and defines a type of RST (Embry, 2009). The top of the sequence is a prominent disconformity truncating altered ash-filled burrows (Fig. 4.9C).

Sequence VI (L’Orignal – Rockland - Hull formations)

Previous work (Raymond, 1914; Kay, 1937; Wilson, 1946; Uyeno, 1974) interpreted a regional superposition of the Rockland, then Hull formations. Recent work (Oruche et al., 2018, 2019) recognized a previously unknown lowermost division (Unit H1, Fig. 4.2) of the Hull Formation requiring re-interpretation of this regional architecture. Representing the once lower part of the Rockland Formation is the newly defined L’Orignal Formation (Chapter 2), extending across the basin, mostly a few metres in thickness except where it thickens near the intersection of extant faults in the central embayment (Fig. 4.9A). Subsequent platform segmentation created local shallow-water carbonate banks (the
revised Rockland Formation; Oruche et al., 2018) and deeper-water ramp settings (Unit H1, Hull Formation) (Fig. 4.9A). The timing of platform segmentation is defined by the Millbrig bentonite (453.36 ± 0.38 Ma; Oruche et al., 2018) found at one site (Loc. 1, see inset map in Fig. 4.9) at the top of the L’Orignal Formation, and characterizing the Turinian-Chatfieldian boundary (Leslie and Bergström, 1995). The segmented platform paleogeography was eliminated through accumulation of regionally expanding high-energy crinoidal shoal systems of the Hull Formation.

Transgressive Systems Tract

The L’Orignal Formation documents a deepening upward succession (Fig. 4.2; Table 4.1). A quartz arenite locally overlies the basal disconformity and documents seaward progradation of shoreface siliciclastics across the prior eroded Watertown platform (Oruche et al., 2018). Where the arenite is not preserved, lithoclasts of similar lithology occur in basal limestone that documents regional variation among high-energy (including ooid grainstone) to low-energy (lime mudstone) settings. Transgression, therefore, reworked the earlier siliciclastics associated lowered base level. The remainder of the formation consists of a rhythmic succession of thinly interbedded muddy fossiliferous limestone to unfossiliferous lime mudstone and shale representing subtidal conditions (Table 4.1; Oruche et al., 2018). It is not known if the rhythmic stratigraphy arises from coastal transport of siliciclastic fines or incursion of offshore shale in response to base-level variation (Oruche et al., 2018). We characterize the top of the TST corresponding to the formation top.

Regressive Systems Tract

The L’Orignal-Rockland formation boundary is stylolitic whereas the L’Orignal-Hull
boundary varies from abrupt to erosional (Table 4.1; Fig. 4.9D). The Rockland Formation consists of a rhythmic alternation of mostly thickly bedded mud-rich limestone and thin shale (Table 4.1; Raymond 1914; Oruche et al., 2018), representing a normal-marine shallow-subtidal setting episodically inundated by siliciclastic mud. In the type section at Loc. 3, upsection variation in abundance of gravel-sized skeletal-bearing less muddy limestone suggests appearance of shallower higher-energy conditions. Unit H1 of the Hull Formation consists of interbedded coarse-grained skeletal-rich limestone and fossiliferous calcareous shale (Table 4.1, Oruche et al., 2018). Large-scale cross-stratification and lateral accumulation of gently inclined discontinuous limestone beds (over 10s of metres along primary dip) are well developed (Oruche et al., 2018). These attributes suggest deposition along a low-gradient bathymetry, one subject to alternating influx of siliciclastic fines related to bottom-water flow (Oruche et al., 2018, 2019).

The above bipartite stratigraphic division is eliminated with appearance of units H2 and H3 of the Hull Formation (Table 4.1; Fig. 4.9). These correspond to Uyeno’s (1974) lower and upper divisions of this same formation but are not necessarily equally represented at each locality. Unit H2 consists of anastomosing lenses and interbeds of fine-to coarse-grained skeletal (crinoidal) grainstone and shale whereas Unit H3 consists of thick (up to metre-scale) beds of coarse-grained encrinites and interbedded lime mudstone and shale (Table 4.1; Oruche et al., 2018). Syndepositional deformation affects both units, with examples of sedimentary boudinage, and both truncated and enrolled bedding. Units H2 and H3 characterize a depositional continuum of shallowing and lateral growth of km-scale crinoid shoal systems: Unit H2 represents the deeper water
(downslope) sand-sized facies; Unit H3 characterizes the shoal platform top environment, but with lower energy intershoal settings (Uyeno, 1974; Oruche et al. 2018).

We interpret maximum flooding to be positioned at the base of Unit H1 coincident with segmentation of the L’Orignal paleoplatform. The equivalent surface is more difficult to place in relation to the Rockland Formation. As part of a common limestone-shale stratigraphic motif, thinner limestone beds in the L’Orignal (TST) succession compared to the Rockland Formation might reflect lower accumulation rates during transgression than during regression (Crevello, 1991). This is supported by an accompanying shallowing upward succession at Loc. 3 within the Rockland Formation (Fig. 4.9A). The Hull (H1 to H3) succession characterizes shallowing of high-energy encrinite shoals and disconformable overstepping of the Rockland banks (Fig. 4.9A).

**Sequence VII (Verulam-lower Lindsay formations)**

**Transgressive Systems Tract**

The base of the TST is marked by abrupt deepening (Kiernan, 1999) across the Hull shoal deposits and marks a seaward shift in depositional locus from the shallow inner to deeper mid- to outer-ramp settings. The TST consists of thinly bedded redepotited skeletal and carbonate lithoclastic sediment interbedded with calcareous shale, and associated with sedimentary structures (graded bedding, hummocky cross-stratification (Fig. 4.10B), and scour features) that illustrate storm activity (Kiernan, 1999). An upsection increase in shale abundance occurs midway through the formation (Fig. 4.10A) and appears to coincide with abrupt deepening along the distal platform-margin in the western Quebec
Fig. 4.10: Lithic attributes of the Verulam to Lindsay formations, Ottawa Group. (A) Generalized lithostratigraphic core section from Loc. 5 with disconformable contact separating the Ottawa Group from Billings Formation. D-S patterns and related symbols are explained in Fig. 4.2B. (B) Outcrop exposure (Loc. 36) of alternating limestone-calcareous shale stratigraphy of the Verulam Formation with hummocky cross-stratification. Coin (23.81 mm diameter) for scale. (C) Outcrop exposure (Loc. 36) of well-bedded Verulam limestone (a) grading into overlying nodular Lindsay limestone (b). (D) Outcrop exposure (Loc. 37) illustrating the transition from nodular shale-lime mudstone fabric (a) of the lower Eastview Member into interbedded organic-rich shale and lime mudstone (b) of the upper Eastview Member. Metre stick (middle ground) for scale. (E) Core section (Loc. 5) illustrating the same planar interbedding of lime mudstone (a) and fossiliferous shale (b) forming the upper Eastview Member.
Basin and shallowing within the more craton interior in central Ontario (Dix and Al Dulami, 2011).

**Regressive Systems Tract**

The Lindsay-Verulam formation succession defines gradation from an interbedded shale-carbonate fabric described above into bioturbated nodular grainstone to packstone of the Nepean Point Member of the Lindsay Formation (Fig. 4.10B and C; Kiernan, 1999; Gbadeyan and Dix, 2013). This suggests that the maximum flooding surface lies in the uppermost Verulam Formation. The upsection facies change records establishment of a well agitated normal-marine inner-platform setting, with ample sunlight based on preserved calcareous algal fragments (Gbadeyan and Dix, 2013). Periods of substrate erosion and non-deposition are marked by formation of paleosurfaces with 3-dimensional (thallassinid-type) burrow networks and *Skolithos* sp. (Gbadeyan and Dix, 2013).

**Sequence VIII (Eastview Member, Lindsay Formation)**

**Transgressive Systems Tract**

A gradational contact separates the relatively high-energy Nepean Point facies from the sand-sized and more shaley Eastview Member that forms the upper Lindsay Formation (Fig. 4.10A), the final stage of carbonate-platform development in this part of the regional foreland (Fig. 4.2). Accompanying an upsection increase in shale abundance is a pronounced change in depositional fabric from concretionary or nodular fossiliferous lime mudstone to planar interbeds of these same lithologies (Fig. 4.11D, E; Table 4.1). This change suggests the emergence of some rhythmic extrinsic control causing segregation of siliciclastic and carbonate sediments. Shales display variable but, overall,
greater organic richness, and beds to lenses of trilobite- and brachiopod-rich skeletal rudstone often cap lime-mudstone beds (Fig. 4.11E; Sharma et al. 2003; Dix and Jolicoeur, 2011). Fossil-rich beds and lenses may represent hiatal or event (storm) accumulations (Kidwell, 1993), or possibly spikes in biological productivity as suggested for the equivalent Collingwood Shale in the eastern reaches of the Michigan Basin (Brett et al., 2006). The top of the platform is not preserved in outcrop or core. From cores at Locs. C and S (inset map in Fig. 4.12) platform stratigraphy is interpreted to have been truncated by a disconformity with increased (~6 m) stratigraphic omission directed toward the central embayment (Gbadeyan and Dix, 2013).

**Sequence IX (foreland basin)**

The Billings Formation forms the basal unit of the foreland-basin succession (Fig. 4.2). A lower division of organic-rich black claystone documents anaerobic to dysoxic deep-water basin conditions (Sharma et al., 2003). Absence of any deepening indicators suggests that this lower division marks maximum water depth, part of the regional expansion of the foredeep across Laurentia (Lehmann et al., 1995).

*Regressive Systems Tract*

Increased abundance of coarse silt to fine-grained sand laminae to beds that first appear in the lower Billings Formation characterizes distal turbidites related to northward migration of orogen-derived siliciclastics (Sanford, 1993; Sharma et al., 2003). The succeeding Carlsbad Formation (Fig. 4.2) records further shallowing from siliciclastic (siltstone, sandstone) to mixed siliciclastic-skeletal carbonate settings along a bathymetric gradient (Sharma et al., 2003). Further shallowing resulted in accumulation of reddish muddy coastal siliciclastics and rare ooid-bearing limestone (Sharma and Dix, 2004), part
of the lower Queenston Formation (Fig. 4.2; Sharma et al., 2003). Subsequent deposition, as part of the evolving regional deltaic system directed toward the Laurentian interior (Brogly et al., 1998), has not been preserved.

Discussion

Regional and global controls on base-level change

Our analysis reveals different scales of TST-RST cycles and base-level change preserved within the Middle to Upper Ordovician foreland succession of the Ottawa Embayment. A low-order TST-RST cycle characterizes the entire preserved Tippecanoe I Megasequence (Fig. 4.2): the TST consists of net platform deepening from peritidal (Providence Island Formation) to outer ramp (upper Lindsay Formation) conditions, and spans ~15 million years; in contrast the RST, represented by only the early history of orogen-derived basin fill, returned water depths to peritidal conditions within ~ 3 million years of platform foundering. This history defines a tectonic cycle formed in response to, first, diachronous foundering of the regional Appalachian foreland platform toward the craton interior in response to lithosphere loading (Bradley and Kidd, 1991; Lavoie, 1994; Macdonald et al., 2017), then erosion and northward transport of orogen-derived (basin-fill) siliciclastics (Ettensohn, 2008; Lavoie, 2008).

There is a higher-order association between sequence development and both tectonic development of the orogen (Fig. 4.2) and climate change:

(1) Within available geochronology, the Sauk-Tippecanoe I boundary coincides with culmination of the closure of the Iapetus Ocean basin (Fig. 4.2) and onset of foreland
development along the distal St. Lawrence Promontory (Knight et al., 1991; Dix and Al Rodhan, 2006). Such coincidence was attributed by Dix and Al Rodhan (2006) to crustal instability beneath the embayment allied with reactivation of faults that defined the buried intracratonic Neoproterozoic aborted rift as a far-field response to distal plate-boundary deformation. The disconformity at the base of the Providence Island Formation is traced south through New York and Vermont (Landing and Westrop, 2006), and may correlate with a mid-Beekmantown disconformity farther south in Virginia (central-eastern USA). Finney et al. (2007) interpreted this latter disconformity to mark onset of inter-regional regression related to eustatic drawdown, and the base of the Tippecanoe I Megasequence boundary. Subsequently, however, the boundary was placed at the top of the regional Beekmantown paleoplatform given that it is easier to document subsequent cratonic-scale foreland flooding (Berezinski et al., 2012).

(2) The voluminous influx of Rockcliffe siliciclastics appears to follow subduction reversal along the Laurentian plate boundary, and formation of a retroarc foreland (Fig. 4.2; Macdonald et al., 2017). Such reversal has been considered responsible for change in regional depositional polarity extending well into the Laurentian interior (Coakley and Gurnis, 1995; Ettensohn, 2008). Such craton-interior uplift provides a mechanism to promote seaward sediment transport and along with reactivation of basement faults beneath the embayment that could explain local fracturing and structurally-induced channelization that influenced early Rockcliffe sedimentation. Lithic transformation from the tide-influenced carbonate-dominated Providence Island to similar tide-influenced, but siliciclastic-dominated Rockcliffe Formation was created not by a significant change in
depositional environment but by increased siliciclastic supply that clearly exceeded local sea-level rise.

However, there is also need for increased humidity (and surface water) in order to enable sediment transport from the craton interior. Indeed, the Middle Ordovician was a period of climatic transition from a warm (to hot) Early Ordovician (Trotter et al. 2008; Quinton et al., 2018) to cooler latest Ordovician (Patzkowsky et al., 1997; Fortey and Cocks, 2005). Cooling to equatorial temperatures of present-day conditions was established by the late Darriwilian (Vandenbrouke et al., 2009). In this time period, southern Laurentia and adjacent Balto-Scandia had transformed to more humid tropical conditions (Pope and Read, 1998; Kiipli et al. 2017).

The Rockcliffe Formation appears coeval with the extensive St Peter Sandstone in the United States that resulted in an expansive craton-interior Middle Ordovician siliciclastic cover (Fig. 4.11; Dake, 1921; Dapples, 1955; Dott et al., 1986). Interior uplift timed with subduction polarity and retroarc development may have initiated the Transcontinental Arch creating a cratonic-scale source area, but with a myriad of local structural highs and lows (Fig. 4.11; Carlson, 1999), enabling fluvial transport to the subtidal regime with redistribution by waves and tides.

(3) The most rapid phase of foreland-platform deepening within the Ottawa Embayment spanned ~8 million years through accumulation of the Ottawa Group (Fig. 4.2). Net eustatic sea-level rise from the Middle through Late Ordovician (Haq and Schutter, 2008; Zhang, 2011) ensured sustained marine conditions over the long term. Of the three higher-order stages of Late Ordovician flooding defined by Zhang (2011), two were
Fig. 11: Geographic distribution of Middle Ordovician siliciclastic sedimentary cover in central to central-eastern North America: Rockcliffe Formation (Ottawa Embayment) and seaward correlation with the Chazyan reefal succession; and, the more expansive St. Peter sandstone over much of the central USA (based on Dake, 1921; Dapples, 1955; and Dott et al., 1992). The horst-graben pattern along the Transcontinental Arch is based on Carlson (1999). Occurrences of outlier Middle Ordovician sandstones are based on Kang (2018) and Dapples (1955). Paleoenvironment abbreviations: f-a, fluvial > aeolian dominance; a-f, aeolian > fluvial.
likely structurally driven and the third is not recorded in the embayment. First, evidence for increased flooding by the early Chatfieldian (Zhang, 2011) marks the transgressive phase of Sequence VI and is coincident with platform segmentation into banks and deeper ramp settings (Oruche et al., 2018). This local history fits with the hypothesis that Chatfieldian flooding along southern Laurentian platform was related to plate-boundary loading (Ettensohn, 2008). In the embayment, platform segmentation and rapid rise in sea level across the Turinian-Chatfieldian boundary was likely structurally induced. Second, a subsequent Edenian flooding phase (Zhang, 2011) appears in the embayment in the form of deepening through the Eastview Member (TST, Sequence VIII) of the Lindsay Formation. But this event also corresponds to platform foundering and local micrograben development across the top of the paleoplatform (Dix and Joliceour, 2011). These events appear timed with emplacement of the Taconic allochthon in the New England portion of the orogen (Fig. 4.2) and similar structural events along strike in the northern Appalachian Basin (Jacobi and Mitchell, 2002). Subsidence and local structural control rather than eustatic rise may have been more influential in driving the Edenian record of flooding in the embayment. Third, the final Richmondian phase of flooding is not recorded in the embayment because this region lay within the perimeter of foreland basin fill related to development of the regional clastic wedge.

Both Chatfieldian and Edenian periods of directed deepening may be better associated with phases of retro-arc shortening during the Late Ordovician (Fig. 4.2; Macdonald et al., 2017). Tectonic shortening across the foreland involves thrust-load stacking within the orogen and regional flexure such that the foreland interior undergoes initial uplift, then foundering (DeCelles, 2012). Development of sustained peritidal conditions
extending across the embayment and extending into the western Quebec Basin represented by Pamela Formation (Harland and Pickerill, 1981; Salad Hersi and Dix, 1999) may record platform-interior flexure.

**Shorter-Term Intrabasinal Controls**

Our analysis demonstrates that the majority of formation boundaries with the foreland succession of the Ottawa Embayment correspond to sequence boundaries (Fig. 4.4). Wilson’s (1946) succession of beds (now formations) of what is now the Ottawa Group (Uyeno, 1974) was defined on the basis of biotic-assemblage succession. We suggest that such biotic change, when placed in context of Walther’s Law, provides evidence for episodic deposition of disconformable (or otherwise abruptly superposed) depositional sequences related to abrupt shifts in depositional settings across a ramp geometry.

We interpret the most prominent examples of abrupt shifts to reflect intrabasinal tectonic influence on base-level change and sedimentation patterns. First, the abrupt development of shale-basin successions in the lower parts of sequences II and III on estuarine and restricted carbonate-platform environments, respectively. Structural influence best accounts for the subsequent heterogeneous lithostratigraphy, especially thick sandstones, within these shale-basin successions through differential subsidence (block faults) of the prior paleoplatform and formation of irregular paleotopography (Dix et al., 2013). Second, the abrupt appearance of middle-ramp facies of Sequence VII on inner-platform shoal facies of the Hull Formation (upper Sequence VI) may be the initial stages of subsequent developed tilting of the Verulam platform in response to distal platform-margin drowning (Dix and Al Dulami, 2011). And, third, differential erosion across the
top of the Ottawa Group (and foreland platform) is best described through local to regional structural controls (as discussed above).

There is evidence for differential subsidence across the embayment during the Turinian as defined by pronounced lateral variation in formation thicknesses (Fig. 4.9A; Salad Hersi and Dix, 1999; Oruche et al., 2018). The location of greater thickness anomalies during this period remains within the central embayment, along the intersection of present-day regional Gloucester and Rideau faults (Fig. 4.1). Indeed, chemostratigraphic ($\delta^{13}$C) evidence suggests abrupt changes in oceanographic patterns and sedimentation rates coincide with formation thickness patterns. Paleostress-field analysis suggests that the extant faults preserve evidence of reactivation timed to plate-boundary orogenesis (Rimando and Benn, 2005); stratigraphic evidence strongly suggests sustained crustal instability throughout the Ordovician foreland in response to distal tectonic drivers (Oruche et al., 2018).

The geometry of an epicontinental carbonate ramp relative to falling sea level predicts coastal encroachment of sandstone, if available (Fig. 4.3A). High-order stratigraphy of sandstone beds (thus excluding the Rockcliffe Formation) is restricted entirely to within the Whiterockian through Turinian platform succession (Fig. 4.2). Appearance of sandstone rather than only a disconformity truncating the carbonate platform may correspond to more significant drops in base-level or, if lateral variation in distribution is demonstrated (such as for the Pamelia Formation; Fig. 4.7), development of depositional conduits allowing sandstone bypassing of the inner platform. Whether such variation correspond to known high-order eustatic variation (Leslie and Lehnert, 2005; Haq and Schutter, 2008) requires improved biostratigraphic control for the embayment. However,
we can begin to address the role of eustatic variation on this time scale through comparison with sequence frameworks of the northern Appalachian Basin.

**Comparison of Sequence Frameworks: Appalachian Basin and Ottawa Embayment**

Sequence notation for the Appalachian Basin is based on Mohawkian (M) and Cincinnatian (C) divisions defined from the central region by Holland and Patzkowsky (1996, 2008) (Fig. 4.1A). From a transect south of the Adirondack Highlands, Joy et al. (2000) interpreted tectonic factors having modified eustatic signals. Correlation of two widely separated sections (New York, Kentucky), led Brett et al. (2004) to interpret regional commonality in stratigraphy underscoring a eustatic framework. Sell et al. (2015) revised Brett et al.’s (2004) sequence divisions using volcanic ash-bed correlation across the eastern and central parts of the USA (Fig. 4.1A). Our comparison is focused on the succession equivalent to the Lowville through Lindsay formations for which there is greater biostratigraphic and geochronological controls. There is insufficient age control to attempt a correlation of the older Whiterockian sequences of Holland and Patzkowsky (1996, 2008) with the Providence Island through Hog’s Back formations.

**Turinian**

Each of the three sequence frameworks (Fig. 4.12) recognizes the lower Lowville Formation as a transgressive unit extending across the eroded Pamelia (and equivalent) paleoplatform. Such inter-regional uniformity may characterize either a eustatic signature or regional tectonic influence timed with onset of basin deepening. Above this stratigraphic level, however, significant differences emerge with respect to nature and
Fig. 4.12: Comparison of sequence stratigraphic framework of the Ottawa Embayment with frameworks defined for the Appalachian Basin. The southern Ontario section (this study) is based on Oruche et al. (2018). Regional frameworks (A, B) extending across the Appalachian Basin (United States) are from Brett et al. (2004) and Sell et al. (2015), respectively, and utilize the sequence nomenclature of Holland and Patzkowsky (1996). The final framework (C) is based sections through northern New York State (Joy et al., 2000), and formational stratigraphy for this region is shown for comparison with that of the Ottawa Embayment. Distributions of sections forming the basis of sequence frameworks for the Appalachian Basin (United States) are illustrated in Fig. 4.1A. Abbreviations: Eden., Edenian; Mays., Maysvillian; Rich., Richmondian.
position of sequence boundaries (Fig. 4.12) that are not a result of high-order division or generalization of facies successions. The prominent erosional surface that truncates mounds and biostromes forming the uppermost Lowville Formation in the embayment (Fig. 4.9B) is not recognized along the Ontario limits of the northern Appalachian Basin. Instead, small to large microbial mounds with intermound *Tetradium* floatstone define a maximum flooding surface subsequently enveloped by regressive muddy Watertown carbonate, similar to the W2 facies association in the Ottawa Embayment (Oruche et al., 2018). Brett et al. (2004) interpreted a similar TST-RST succession for the Lowville Formation through the Appalachian Basin but Sell et al.’s (2015) re-interpretation differs (Fig. 4.12).

Differential subsidence within the Watertown Formation in the embayment (Fig. 4.9A; see Oruche et al., 2018, 2019) is not recognized in the northern Appalachian Basin. Furthermore, siliciclastics are not reported along the top of the Watertown Formation in New York State that otherwise cap Sequence V in the Ottawa Embayment (Fig. 4.2 and 4.12). By itself, this latter difference could arise from either (1) preferred uplift to the north or (2) unequal advance of siliciclastics during eustatic drawdown. The first possibility could certainly initiate preferential south-directed influx of siliciclastics across the embayment while subtidal conditions were maintained in the northern Appalachian Basin. But in this latter region, a hardground in the Ontario section defines the sequence boundary (MacFarlane, 1992; Oruche et al., 2018) whereas sustained deposition farther south is interpreted to be related to transgression (Brett et al., 2004).

This regional gradient is enhanced by differential distribution of prominent transgressive deposits: in the embayment, they form the lower part of Turinian sequences V and VI in
the embayment. In the northern Appalachian Basin (Ontario), transgressive deposits cap the Watertown Formation (Sequence B) and a differentially eroded Selby Formation (Sequence C) (Oruche et al., 2018). The equivalent boundary to the latter in New York State has been interpreted as being part of continuous transgression (Fig. 4.12; Brett et al., 2004; Sell et al., 2015) and regression (Joy et al., 2000).

Post-Turinian

Platform segmentation in the Ottawa Embayment follows accumulation of the Millbrig Bentonite defining the beginning of the Chatfieldian. In the embayment, this boundary marks the top of the TST of Sequence VI. In the Appalachian Basin, the equivalent paleosurfaces are all sequence boundaries, yet reflect a diversity of processes: erosion, hardgrounds, hiatal surfaces (Fig. 4.12; Joy et al., 2000; Brett et al., 2004; Sell et al., 2015). This highlights a mosaic of depositional conditions related to local variation in bathymetry, oceanography, and productivity. As noted above, timing of platform segmentation in the Ottawa Embayment may be a better fit with regional tectonic control (Ettensohn, 2008) than eustatic rise.

Above this stratigraphic level, regional correlation of Chatfieldian sequences is problematic because of an absence of sufficient biostratigraphic zonation within the embayment. However, three changes in base level can be mapped regionally. First, the prominent base-level rise across the Hull (H3)-Verulam (or sequence VI-VII) boundary in the embayment matches with a slightly older facies shift separating the Kings Falls-Sugar River succession in the northern Appalachian Basin (Fig. 4.12). Such an age difference fits a predicted diachroneity of either foreland flexure or eustatic rise.
extending into the craton interior. Second, base-level fall defining the Verulam-lower Lindsay Formation transition (within Sequence VII) also appears to be in common with the transition from deeper water nodular calcilutites of the Denley Formation up into grainstones of the Rust Formation (Brett et al., 2002). This might mark a eustatic drawdown but could also reflect decrease in subsidence rates. Another common change, but one of structural origin, is structural modification of the foreland platform and development of a prominent foreland disconformity during final phase of platform subsidence (Jacobi and Mitchell, 2002; Dix and Joliceour, 2011; Gbadeyan and Dix, 2013).

Although there remains much uncertainty, we note that all but one of the structurally defined sequence boundaries of Joy et al. (2000) for the northern Appalachian Basin match with sequence boundaries in the Ottawa Embayment (Fig. 4.12). While eustatic rise produced regional onlap extending deep into the craton interior through the Middle to Late Ordovician (Zhang, 2011), we interpret inter-regional differences of sequence positions and contrast in nature of their boundaries to underscore local influence of tectonic frameworks modifying the eustatic signal.

Conclusions

(1) A sequence stratigraphic framework of the Middle to Upper Ordovician foreland succession in the Ottawa Embayment consists of eight transgressive-regressive depositional sequences associated with foreland platform development, and the lower preserved part of a regressive sequence that documents history of subsequent foreland basin-fill following platform foundering.
(2) Subordinate stages of sequence development coincide with stages of orogen development; namely: a) the Sauk-Tippecanoe I sequence boundary corresponds with closure of the Iapetus Ocean Basin; b) voluminous influx of Middle Ordovician siliciclastics coincides with subduction reversal and predicted craton-hinterland uplift; c) deepening from peritidal to outer ramp platform settings occurs over ~8 million years and coincides with retro-arc shortening; and d) structural collapse of the foreland platform and subsequent basin fill coincides with thrust-loading along the plate boundary.

(3) Although hinterland uplift is required to accommodate voluminous influx of Middle Ordovician siliciclastics, their transport was aided by global climate change to more humid temperate conditions.

(4) Most formation boundaries coincide with sequence boundaries, and abrupt changes in base level and paleoenvironmental conditions also coincide with local evidence for synsedimentary structural deformation and larger scale intrabasinal differential platform subsidence. Structural control within the embayment arises from reactivation of inherited basement faults associated with a shallowly buried Neoproterozoic intracratonic graben as a far-field response to plate-boundary tectonism.

(5) Comparison of sequence stratigraphic frameworks for the Ottawa Embayment and northern Appalachian Basin reveals mostly conflicting positions and character of boundaries with frameworks proposing eustatic control. In contrast, there is greater compatibility with the sequence framework proposed by Joy et al. (2000) defining greater tectonic control.
CHAPTER 5: CONCLUSIONS

This study examined the Middle to Upper Ordovician foreland succession of the Ottawa Embayment, and in particular the Upper Ordovician Turinian-Chatfieldian interval. The purpose was to utilize a more modern integrated stratigraphic approach to characterize and correlate depositional and stratigraphic architectures with equivalent successions in the adjacent Quebec Basin to the east and Appalachian Basin to the south; and, in doing so, draw out eustatic, tectonic, and climatic controls. The role of tectonism is of particular interest because the embayment falls along the same axis as a Neoproterozoic intracratonic fault system, now manifest as the Ottawa-Bonnechere graben. The following conclusions are grouped according to organization of individual chapters (2-4) in this thesis.

Upper Turinian – lower Chatfieldian (Upper Ordovician) stratigraphy in the embayment preserves three stages of carbonate-platform development. The first and last represent local expressions of apparent inter-regional development of depositional systems suggesting that they reflect either eustatic controls or common inter-regional tectonism. However, when compared to adjacent basins, the intermediate stage contains stratigraphic and depositional attributes both specific to the embayment and some in common with the Appalachian Basin to the south. Specific to the embayment, in ascending order through this succession is: 1) a transgressive deposit generated during westward shoreline translation that defines the base of the Watertown Formation, and dramatic east-directed thickening of the formation; 2) post-Watertown erosion and progradation of shoreface siliciclastics; and, 3) platform segmentation immediately following the Turinian-Chatfieldian boundary (as defined by local preservation of the Millbrig Bentonite dated at
453.36 ± 0.38 Ma) that produced a paleogeographic mosaic of low-energy carbonate banks and deeper intraplatform settings. The latter event distinguishes a revised Rockland Formation from a hitherto unrecognized lowermost division of the Hull Formation. These embayment-specific events record the influence of syndepositional tectonism, and specifically platform segmentation appears coincidental with interpreted increase in regional subsidence across southern Laurentian and changing oceanographic conditions. In common with the Appalachian Basin, however, is a significant change in lithofacies across the Turinian-Chatfielidian boundary that, to the south, has been linked to regional paleoceanographic change related to increased subsidence of the southern Laurentian platform.

Use of δ¹³C profiles have enabled increased resolution of stratigraphic correlation and change in depositional conditions within the embayment. Four δ¹³C excursions (E1 to E4) are correlative through the upper Turinian-lower Chatfieldian interval along the axis of the Ottawa Embayment and through outliers to the northwest. Excursion E1 is restricted to the Watertown Formation; E2 occurs within the L’Orignal Formation (a newly defined formation, but forming the lower part of the previous workers’ Rockland Formation); E3 is correlated through coeval facies of Rockland Formation and the lowermost Hull Formation (Unit H1), and forms the local expression of the regional (if not global) Guttenberg δ¹³C excursion (GICE); and, Excursion E4 occurs within the upper Hull strata (Unit H3). There is some evidence for diagenesis having modified the magnitude of isotopic expression, especially beneath disconformities. Lateral variation in the magnitude of δ¹³C excursions through Ottawa Embayment reflect local modulation of regional to global events through influence of syndepositional structural control that
influenced local variation in oceanographic exchange, water depth, accumulation rates, and productivity.

In order to better understand the Turinian-Chatfieldian succession, a sequence stratigraphic framework was established for the entire Middle to Upper Ordovician foreland succession (equivalent to the Tippecanoe I Megasequence) in the Ottawa Embayment. This study re-affirms placement of the Sauk-Tippecanoe I megasequence boundary within, and not at the top of, the Beekmantown Group, at the Fort Cassin-Providence Island boundary. A tectonic cycle of, first, foreland-platform deepening, then basin fill documents a first-order transgressive-regressive sequence. Eight higher order T-R cycles comprise the platform succession, and a regressive succession represents the basin-fill history. The majority of these depositional sequences can be categorized as being of tectonic origin, linked in time with orogenic events or intrabasinal evidence of synsedimentary structural control. Global climate change during the Middle Ordovician resulted in voluminous influx of siliciclastics that define the Rockcliffe Formation, part of a regional siliciclastic sedimentary cover across the eastern and central United States. When compared to conflicting sequence frameworks (supporting eustatic versus tectonic controls) published for the Appalachian Basin, there are prominent differences in the position and nature of depositional sequence boundaries of eustatic frameworks. Instead, there appears to be greater compatibility with a tectonic framework.

Although evidence for tectonic control within a foreland basin should not be surprising, this study has defined an apparent elevated record of syndepositional structural control within the Ottawa Embayment when compared to the Appalachian Basin to the south. This greater sensitivity likely arises due to the inheritance, and response to regional
tectonism, of a structurally weakened continental crust beneath the Ottawa Embayment defined by the axis of a Neoproterozoic intracratic fault system. Tectonism is viewed as being important in influencing higher order base-level change superimposed on net eustatic rise through the Middle and Late Ordovician.
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## Appendix A: Section locations, geographic coordinates and related information

<table>
<thead>
<tr>
<th>Locality</th>
<th>Name</th>
<th>Latitude* (° N)</th>
<th>Longitude* (° W)</th>
<th>Type**</th>
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<td>Rockcliffe Parkway</td>
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</table>

* derived from Google Earth (2016)

** c, core; o, outcrop, q, quarry: operational (o), non-operational (n)
Appendix B: Methodology for XRD, U-Pb geochronology, δ^{13}C and δ^{18}O

X-ray Diffraction

The mineralogy of bulk materials and clay-size separates was determined by X-ray powder diffraction analysis (XRD) at the Geological Survey of Canada (Ottawa). Bulk samples were micronized using a McCrone mill in isopropyl alcohol or distilled water until a grain size of about 5-10 µm was obtained. The samples are dried and then back pressed into an aluminum holder to produce a randomly oriented specimen. For clay-size separates, 40 mg was suspended in distilled water and pipetted onto glass slides and air-dried overnight to produce oriented mounts. X-ray patterns of the pressed powders or air-dried samples were recorded on a Bruker D8 Advance Powder Diffractometer equipped with a Lynx-Eye Detector, Co Kα radiation set at 40 kV and 40 mA. Samples were scanned from 2-86° 2-theta. The samples were also analyzed following saturation with ethylene glycol, and scanned over the same 2-theta range, and following heat treatment (550°C for 2 hours) but scanned from 2 to 35° 2-theta. Identification of mineralogy was made using EVA (Bruker AXS Inc.) software with comparison to reference mineral patterns using Powder Diffraction Files (PDF) of the International Centre for Diffraction Data (ICDD) and other available databases.

U-Pb CA-ID-TIMS Geochronology

U-Pb isotopic data were obtained at the Jack Satterly Geochronology Laboratory at the University of Toronto. Zircon grains were pre-treated by chemical abrasion to remove radiation-damaged and altered zones (Mattinson, 2005). Grains were placed in a muffle furnace at ~1000°C for ~48 hours to restore crystallinity to radiation damaged zones, and
then partially dissolved in 50% HF in Teflon dissolution vessels at 200°C for approximately 17 hours. Each zircon was cleaned in HNO\textsubscript{3} at room temperature, and transferred to a miniaturized Teflon bomb (Krogh, 1973). A mixed \textsuperscript{205}Pb-\textsuperscript{233}U-\textsuperscript{235}U spike (EARTHTIME community tracer ET535, to facilitate inter-laboratory comparisons, see www.earth-time.org) was also added to the Teflon dissolution capsules. Zircon was dissolved using \textasciitilde0.10 ml concentrated HF acid and \textasciitilde0.02 ml 7N HNO\textsubscript{3} at 200°C for 5 days, then dried to a precipitate and re-dissolved in \textasciitilde0.15 ml of 3N HCl. Uranium and lead were isolated from the zircon solutions using anion exchange resin, evaporated in dilute phosphoric acid, and deposited onto out gassed rhenium filaments with silica gel to stabilize emission (Gerstenberger and Haase, 1997). Samples were analyzed with a VG354 mass spectrometer using a Daly detector in pulse counting mode for Pb; U was measured either in static mode using an array of 3 Faraday collectors or in dynamic mode using a single Daly detector. Corrections to the \textsuperscript{206}Pb-\textsuperscript{238}U and \textsuperscript{207}Pb/\textsuperscript{206}Pb ages for initial 230Th disequilibrium in the isotopic data have been made assuming a Th/U ratio in the magma of 4.2. Laboratory procedural blanks are routinely at the 0.5 pg and 0.1 pg level for Pb and U, respectively. All common Pb was assigned to procedural Pb blank. Dead time of the ion counting system for Pb was 16 ns and for U was 14 ns. The mass discrimination correction of the Daly detector is constant at 0.05% per atomic mass unit. Amplifier gains and Daly characteristics were monitored using the SRM 982 Pb standard. Thermal mass discrimination correction for Pb is 0.10 % per atomic mass unit. U fractionation was measured internally and corrected for each cycle in the static measurements. Decay constants (\textsuperscript{238}U and \textsuperscript{235}U are 1.55125 X 10-10/yr. and 9.8485 X 10-10/yr., respectively) are from Jaffey et al. (1971). Age calculations performed using an
in-house program by D.W. Davis. All age errors quoted in the text, Appendix E, and error ellipses in the Concordia diagram (Fig. 2.9D), and error bars on the mean age plot are given at the 95% confidence interval. Plotting and age calculations were done using Isoplot 3.00 (Ludwig, 2003).

δ¹³C and δ¹⁸O

At the Queen's Facility for Isotope Research (Queen's University, Kingston, Canada), the δ¹⁸O and δ¹³C ratios of calcite were determined by reacting approximately 1 mg of powdered material with 100% anhydrous phosphoric acid at 72°C for 4 hours. The CO₂ released was analyzed using a Thermo-Finnigan Gas Bench coupled to a Thermo-Finnigan DeltaPlus XP Continuous-Flow Isotope-Ratio Mass Spectrometer (CF-IRMS). δ¹⁸O and δ¹³C values are reported using the delta (δ) notation in permil (‰), relative to Vienna Pee Dee Belemnite (VPDB) and Vienna Standard Mean Ocean Water (VSMOW) respectively, with precisions of 0.2‰.

At the Jan Veizer Stable Isotope Laboratory (University of Ottawa, Ottawa, Canada). Samples are weighed exetainer vial is filled with about 10mL of specially prepared anhydrous phosphoric acid (method adapted from: Coplen et al., 1983), capped, heated and evacuated on a vacuum line for at least 1 hour to degas the acid. Meanwhile about 0.5-0.7mg of sample or standard is measured into clean exetainers. A calibrated internal standard is also loaded with each batch to be run as an unknown. Once the samples and standards have been weighed, the vials are loaded 8 at a time without caps into an extra rack turned on its side (i.e. Exetainers are kept almost horizontal). A 1mL disposable syringe is used to carefully drop 0.1mL of acid just past the threaded top of each
Exetainer. Vials are kept horizontal while being recapped (after every 8). Once all the vials have acid added and caps on, each column of 8 vials is flushed and filled with UHP helium off-line for 4 minutes at a rate of 60-70 mL/min. Prepared vials are then tipped upright and immediately placed in the heated block of the GasBench (either 25.0°C (calcite) or 50°C (dolomite)) and left to react for 24 hours, is followed by extraction in continuous flow. The measurements are performed on a Delta XP and a Gas Bench II, both from Thermo-Finnigan; see Application Flash Report G 31 from Thermo-Finnigan. The analytical precision (2 sigma) is +/- 0.15 permil. Note: data for carbon and oxygen were normalized using international standards NBS-18, NBS-19, and LSVEC (lithium carbonate).
Appendix C: Field description of the stratotype of the L’Orignal Formation, Loc. 1.

The section is exposed along the quarry road down into the bottom of the quarry. The formation occurs 7.7 to 5.8 m beneath the prominent quarry bench. The following records the general lithology from base to top.

<table>
<thead>
<tr>
<th>Lithostratigraphic Unit</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Watertown Formation</td>
<td>As of Summer 2017, the base of the quarry remained within the Watertown Formation, with an exposure of ~ 11 m consisting of mostly thick bedded skeletal mudstone and wackestone.</td>
</tr>
<tr>
<td>L’Orignal Formation</td>
<td>Interbeds of medium-bedded brownish gray, skeletal packstone (90%) and very thin beds of fossiliferous dark gray shale. Along strike the latter vary with shale partings due to a more amalgamated limestone bedding fabric. Benthic fossils include skeletal remains of crinoids (e.g., ossicles), trilobites, disarticulated shells of bivalves, brachiopods (punctate and impunctate) and ostracodes; abundant fragmented to whole bryozoans.</td>
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<tr>
<td>0 – 1.5 m</td>
<td>As above, including fossil remains, but with greater percentage (20%) of shale.</td>
</tr>
<tr>
<td>1.5 – 1.8 m</td>
<td>~10 cm recessive interval of shale and a clay bed, with no obvious macrofossils, and an internal lithostratigraphy as follows (from base to top):</td>
</tr>
<tr>
<td>0 – 2 cm</td>
<td>fissile dark gray siltstone; abrupt upper contact with,</td>
</tr>
<tr>
<td>2 – 7 cm</td>
<td>light grey to white sticky clay, with abundant rounded to euhedral, lustrous mica (biotite) crystals; gradational contact with</td>
</tr>
<tr>
<td>7 – 8 cm</td>
<td>sticky and silty (but non-biotite-bearing) light gray clay; abrupt contact with</td>
</tr>
<tr>
<td>8 – 10 cm</td>
<td>dark grey siltstone.</td>
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<tr>
<td>Hull Formation (Unit H1)</td>
<td>Thin to medium-bedded skeletal-rich grainstone bedded with bryozoan-rich calcareous shale. The basal contact is disconformable. This lithofacies association extends up to the first quarry bench.</td>
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Appendix D: Litho- and chemostratigraphic sections

### Legends

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Loc. 1

samples | lithology
---|---

Unit H3

\[ \delta^{13}C_{VPDB} (\text{‰}) \]

0 | -2

\[ \delta^{18}O_{VPDB} (\text{‰}) \]

-3 | -5
### Loc. 3

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Loc. 5 continuation

lithology

350
355
360
365
370
375
380
385

Unit H3

Loc. 6

samples

lithology

unit H1

BDA16a
BDA5a
5a,7a

BDA1a,2a
3a,4a,8a
10a,11,12
11b

BDA1,1b
2,3,4,5,6
7,8,9,10

Lowville

Water

\[ \delta^{13}C_{VPDB} (\%o) \]

\[ \delta^{18}O_{VPDB} (\%o) \]

-2 0 2

-4 -8 -6
Loc. 6 continuation

[Diagram with labels and data points for unit H3 and H1, showing various categories such as Mudst, Dolom, Waquest, Packst, Granst, Floatst, and Rudst.]
Loc. 7 west

![Diagram of lithology and δ\(^{13}\)C\(_{VPDB}\) and δ\(^{18}\)O\(_{VPDB}\) graphs]
Loc. 8
Loc. 10
Loc. 11

samples

lithology

3
2
1

PRc-5
PRc-4
PRc-1,2,3

Lowville

Mud
dolm
Wackest
Packst
Granst
Floest
Rudst

metres
Loc. 13

(samples) Napanee

(composition) Selby

(composition) Watertown

\[ \delta^{13}C_{\text{VPDB}} \%o \]
\[ \delta^{18}O_{\text{VPDB}} \%o \]

covered
Loc. 14

![Diagram showing samples and lithology](image)
Loc. 15
Loc. 16a
Loc. 16b

samples

lithology

5 metres

TK-h
TK-e,f
TK-d

Watertown

Lowville

Mudst
dolom
Wackest
Packst
Grainst
Floatst
Rudst
Loc. 17

lithology

Glenburnie beds

1

2

3

4

meters

Wetston
dolost
Wackest
Packst
Grainst
Floatst
Rudst
Loc. 22

![Graph showing lithology and isotopic data](image-url)
Loc. 24b
Loc. 26

lithology

Rockcliffe

Providence Island

no measured

Fort Cassin

Metres

day silt vf f ml vc
Loc. 27

samples

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metres

lithology

- Mudst
- silt
- wackest
- packst
- grainst
- floatst
- Rudst

clay
silt
vf
f
m
c
vc
Loc. 34

samples | lithology
-------|--------
RD-3   |       
RD-2,5 |       
RD-4   |       
RD-1   |       

Pamella

Mudst dolom | Wackest | Packst | Grains | Float | Rudent | dolomudstone

mudstone
Appendix E: U-Pb isotopic data for single, chemically abraded zircon grains from bentonite sample LOQ-B, L'Orignal Quarry, Ontario

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<th>No.</th>
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<th>Th/U (pg)</th>
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<th>207Pb/235U</th>
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<th>Error Corr</th>
<th>207Pb/206Pb</th>
<th>± 2σ</th>
<th>Ages (Ma)</th>
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Notes:
Z is zircon; Weight (estimated) is in micrograms.
Zircon grains have been chemically abraded (Mattinson 2005).
Th/U is calculated from radiogenic $^{208}\text{Pb}/^{206}\text{Pb}$ ratio and $^{207}\text{Pb}/^{206}\text{Pb}$ age assuming concordance.
PbC = total common Pb (in picograms); assigned to the isotopic composition of laboratory blank ($^{206}\text{Pb}/^{204}\text{Pb}=18.49±0.4\%; ^{207}\text{Pb}/^{204}\text{Pb}=15.59±0.4\%;$$^{208}\text{Pb}/^{204}\text{Pb}=39.36±0.4\%)$.
$^{206}\text{Pb}/^{204}\text{Pb}$ is corrected for fractionation and common Pb in the spike.
Pb/U ratios are corrected for fractionation, common Pb in the spike, and blank.
Error Corr = correlation coefficients of X-Y errors on the concordia plot.
Correction for $^{230}\text{Th}$ disequilibrium in $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ assuming Th/U of 4.2 in the magma.
Decay constants (Jaffey et al., 1971): $^{238}\text{U}$ and $^{235}\text{U}$ are 1.55125 X 10-10/yr and 9.8485 X 10-10/yr.
Appendix F: Carbon and oxygen isotope data by locality

Isotope data (relative to VPDB) is referred to elevation (m, metres) above the base of an outcrop/quarry section, or to depth (feet, metres) in core.

**Loc.1**

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<th>$\delta^{13}$C 3-pt avg</th>
<th>$\delta^{18}$ O‰ (VPDB)</th>
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Duplicates run but not reported by the laboratory
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![Raw Data vs 3-Pt Moving Average](image-url)
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### Raw Data vs 3Pt Moving Average

![Graph showing raw data vs 3Pt moving average for δ\(^{13}\)C‰ vs VPDB and δ\(^{13}\)O‰ vs VPDB.](image)
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The individual sections arise due to sampling different parts of the quarry that include faults, and establishing correlation for this composite section.

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![Chart Title](chart.png)

- \( \delta^{13}C \% \text{ VDPB} \)
- \( \delta^{13}C \% \text{ 3-Pt Moving Avg} \)
**Loc. 21**

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![Raw Data vs 3-Pt Moving Average](chart.png)
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Duplicate Analysis

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Appendix G: Carbon and oxygen stable isotope section profiles and cross-plots

Profiles (left diagram in each row) are shown measured (metres) from base of section, or relative to footage in drill core (Locs. 4 and 5). Cross-plots (right diagram in each row) also include Pearson correlation coefficient (represented by $r^2$) and projected regression line (blue). Variables are shown for only the top diagrams on each page. Plots were generated using Deltagraph Professional. Double arrows identify horizons (or intervals) define types of stratigraphic covariance: a, beneath erosional or hardground surfaces; b, enhancing limits of excursions; c, related to transgressive facies; and, d, no obvious association. Lithostratigraphic information is provided in Fig. 3; grey lines define formation boundaries.
Appendix H: Maximum and baseline $\delta^{13}$C values* associated with other sites of the Guttenberg $\delta^{13}$C excursion (GICE), eastern North America**

<table>
<thead>
<tr>
<th>Locality</th>
<th>Formation</th>
<th>$\delta^{13}$CVDPB (%)</th>
<th>Source max</th>
<th>baseline</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Owen Quarry, Ontario/Quebec</td>
<td>unnamed formation</td>
<td>2.2</td>
<td>0.2</td>
</tr>
<tr>
<td>a</td>
<td>northeast- Missouri</td>
<td>Decorah Fm</td>
<td>2.5</td>
<td>-1.0</td>
</tr>
<tr>
<td>b</td>
<td>Harris No.1 drill core in Louisa county, Iowa</td>
<td>Decorah Fm</td>
<td>1.4</td>
<td>0.3</td>
</tr>
<tr>
<td>c</td>
<td>Brook Farm’s SS-9 Drillcore, Iowa</td>
<td>Decorah Fm</td>
<td>1.3</td>
<td>-0.5</td>
</tr>
<tr>
<td>d</td>
<td>Cominco SS-2 Drillcore, Jackson County, Iowa</td>
<td>Decorah Fm</td>
<td>2.3</td>
<td>-0.2</td>
</tr>
<tr>
<td>e</td>
<td>McGregor Quarry, Clayton County, Iowa</td>
<td>Decorah Fm</td>
<td>1.1</td>
<td>-1.0</td>
</tr>
<tr>
<td>f</td>
<td>Cominco SS-12 Drillcore, Jackson County, Iowa</td>
<td>Decorah Fm</td>
<td>2.4</td>
<td>-2.0</td>
</tr>
<tr>
<td>g</td>
<td>Elkader A1 Drillcore, Clayton County, Iowa</td>
<td>Decorah Fm</td>
<td>1.5</td>
<td>-0.6</td>
</tr>
<tr>
<td>h</td>
<td>Big Spring No.4 Drillcore, Clayton County, Iowa</td>
<td>Decorah Fm</td>
<td>1.3</td>
<td>-0.4</td>
</tr>
<tr>
<td>i</td>
<td>Kirkfield Quarry, Lake Simcoe Region, Ontario</td>
<td>Bobcaygeon Fm</td>
<td>1.7</td>
<td>0.9</td>
</tr>
<tr>
<td>j</td>
<td>Great La Cloche Island, Manitoulin, Ontario</td>
<td>Bobcaygeon Fm</td>
<td>2.0</td>
<td>0.2</td>
</tr>
<tr>
<td>k</td>
<td>Roaring Brook, Martinsburg, New York</td>
<td>Napanee Fm</td>
<td>2.6</td>
<td>2.3</td>
</tr>
<tr>
<td>l</td>
<td>Reedsville, Mifflin County, Pennsylvania</td>
<td>Salona Fm</td>
<td>3.3</td>
<td>0.8</td>
</tr>
<tr>
<td>m</td>
<td>Dolly Ridge, Pendleton County, West Virginia</td>
<td>Dolly Ridge Fm</td>
<td>3.0</td>
<td>0.5</td>
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<td>n</td>
<td>Hagan, Lee County, Virginia</td>
<td>Trenton Limestone</td>
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<td>0.5</td>
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<tr>
<td>o</td>
<td>Mnerva Drillcore, Northern Kentucky</td>
<td>Lexington Fm</td>
<td>2.0</td>
<td>0.5</td>
</tr>
<tr>
<td>p</td>
<td>Composite: Boonsborough-Frankfort-Minerra Core</td>
<td>Lexington Fm</td>
<td>2.7</td>
<td>0.5</td>
</tr>
<tr>
<td>q</td>
<td>Composite: Nashville Airport-Bladeville-Tanglewood</td>
<td>Hermitage Fm</td>
<td>2.9</td>
<td>0.4</td>
</tr>
<tr>
<td>r</td>
<td>Anticosti Island (drillcore), Quebec</td>
<td>Mingan Gp</td>
<td>~1 (?)</td>
<td>~1 (?)</td>
</tr>
<tr>
<td>s</td>
<td>Maxon Roger 1 (drill core), New York</td>
<td>Trenton Gp</td>
<td>2.8</td>
<td>?</td>
</tr>
<tr>
<td>t</td>
<td>Smith Frank 1 (drill core), New York</td>
<td>Trenton Gp</td>
<td>2.9</td>
<td>~0.6</td>
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<tr>
<td>u</td>
<td>Olin 1 (drill core), New York</td>
<td>Trenton Gp</td>
<td>2.8</td>
<td>?</td>
</tr>
<tr>
<td>v</td>
<td>Carter 1 (drill core), New York</td>
<td>Trenton Gp</td>
<td>2.5</td>
<td>~0.3</td>
</tr>
<tr>
<td>w</td>
<td>Richards 1 (drill core), New York</td>
<td>Trenton Gp</td>
<td>2.7</td>
<td>~1.3</td>
</tr>
</tbody>
</table>

* a 3-point running average was used similar to the approach in this study.
** the excursions referred to the GICE at Dexter Quarry (New York) and Hull Cement Plant (Gatineau, Quebec) by Barta et al. (2007) are interpreted to be younger events, possibly equivalent to excursion E4 in the Ottawa Embayment.
*** minimum values associated with an interpreted post-peak record.
Appendix I: Trace Fossils and Gastropods, Fort Cassin Formation (Loc. 24 and 25)

Summary by George R. Dix, Department of Earth Sciences, Carleton University

Loc. 24: Ichnofossil Assemblage

Description

Superposition of laminated, sandy and lithoclastic dolostone of the Providence Island Formation on strongly bioturbated dolomudstone of the Fort Cassin Formation establishes a prominent abrupt change in facies. The burrow fabric is erosionally truncated. Rock slabs cut perpendicular to bedding shows two (T1, T2) groups of burrows. T1 represents the pervasive 3-dimensional fabric characterizing excavation of the host sediment and filled with lighter dolomudstone. There is no obvious preferred development of bioturbation, the fabric create a gallery effect with angular burrow boundaries. A possible U-shaped connectivity is locally developed (see Fig. 4.6A) or is coincidental. In contrast to this fabric are T2 burrows that consist mostly of small round to ovoid unlined burrows that occur only in the dolomudstone filling the T1 network. T2 burrows appear nested, each group is organized along a gentle slope with groups oriented parallel to each other. There is an example of an apparent cross-section of a small spiral form of T2 burrows. Along the margins of both burrow systems, the host sediment is darkened (dark grey, black). Petrography and SEM analyses show that these darker areas are not pyritic. Dolomite crystals cross-cut halo boundaries indicating that darkening represent relict (ghost) features of sediment prior to dolomitization.
**Interpretation**

T1 excavations may be referred to Thalassinoides sp., there being similar geometries as noted in both Ordovician and younger strata (Savrda 1991; Sheehan and Schiefelbein, 1984). The apparent spiral fabric of a T2 burrow may refer to Gyrolithes sp., a burrow type that has been noted in combination with Thalassinoides (Bromley, 1990). However, this trace fossil is found usually only in post-Paleozoic strata and only diminutive forms are known from the Cambrian and Upper Paleozoic strata (see references in Netto et al. 2007). Otherwise, unlined T2 burrows are normally referred to Planolites sp. (Buatois and Mángano, 2011). Burrow fill of both the T1 and T2 groups is not related to the lithology of the overlying Carillon Formation. Thus, absence of an equivalent gastropod-lithic rudstone at this site, but caps the Fort Cassin Formation ~50 m adjacent (Loc. 25) suggests differential erosion across the post-Fort Cassin paleosurface between the two sites. There is no need for the rudstone to have extended to Loc. 24 at all. The cross-cutting burrow system exhibited at Loc. 24 defines tiering (Bromley, 1990) that originates through an active response to changing environmental conditions. In this case, transition from a once open-excavation of the seafloor to smaller discrete burrows could arise through decrease in oxygen restricting active vertical-to-horizontal excavation (Bromley, 1990). If there was differential synsedimentary uplift of Loc. 25 relative to Loc. 24, influenced by a syndepositional fault, then the surface of the Fort Cassin platform may have developed as a mosaic of small ponds of restricted (warmer, more saline?) ocean conditions separated by small(er)(?) uplift areas of uplift undergoing vadose diagenesis (see below).
Loc. 25: Gastropod Forms and Paleoenvironmental Significance

Description

Gastropods are preserved in the upper tens of centimetres of the Fort Cassin Formation as dolostone steinkerns, dolomitized shells filled with dolostone, and exposed molds. No whole shell was observed in outcrop. The fossil-rich unit is a gastropod-lithic rudstone: shells are packed and cemented together, and admixed are rounded lithoclasts containing fragments of dolomitized gastropod shells and intraskeletal matrix. This illustrates rapid reworking of material as the unit accumulates.

The gastropods are small, with maximum dimensions of the more complete forms being, typically, < 5 mm. The fossils might represent a micromorphic assemblage. Gastropod cross-sections along bedding planes and vertical exposures, and as seen in thin section, demonstrate a variety of forms (see figure below) that suggest a relatively diverse assemblage.
There are examples of forms with fossilized columnella, others with an umbilicus; and, one example of a whorl that would have been out of contact with its neighbor (Appendix I: Fig. 1j) that may define a ecculiomphalid form. However, as none could be extracted whole, the forms are not definitive of genus or species (written comm., D. Rohr, 2019).
**Interpretation**

Micromorphism arises from a variety of genetic and environmental causes (Hallam, 1965; Mancini, 1978), in particular where physiological and environmental harsh conditions may exist, and where biota represent ecological opportunists following catastrophic collapse of the food chain (Frasier and Bottjer 1991), possibly establishing atypical microbial or algal-foliage supported feeding strategies (Peel, 1978; Arup, 1991). Lithostratigraphy and vadose textures preserved in the gastropod coquina support a shallowing succession to (at times) above the water table. Lithoclasts hosting the same skeletal-matrix fabric as the coquina itself illustrates reworking of shells, some with remnant cement, in a peritidal swash zone. The fossil/sedimentary composition forms an ancient example of the modern Cerithid-rich accumulations in tropical tide pools that are subjected, at times, to extreme tropical mid-day temperatures (Frasier and Bottjer, 2004). Although no algal material is preserved, it may have been non-calcareous although occurrence of a vadose dolomicrite may also identify microbial activity having offered an alternate food-source strategy for gastropods.