Canadian Journal of Earth Sciences
Revue canadienne des sciences de la Terre

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Journal:	<i>Canadian Journal of Earth Sciences</i>
Manuscript ID	cjes-2016-0002.R1
Manuscript Type:	Article
Date Submitted by the Author:	26-Apr-2016
Complete List of Authors:	Zhang, Shunxin; Canada-Nunavut Geoscience Office Mirza, Khusro; Geological consult Barnes, Chris; School for Earth and Ocean Sciences
Keyword:	Upper Ordovician-Upper Silurian, conodont biostratigraphy, Canadian Arctic Islands, Allen Bay and Cape Phillips formations, thermal maturation

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**Upper Ordovician-Upper Silurian conodont biostratigraphy, Devon Island
and southern Ellesmere Island, Canadian Arctic Islands, with implications for
regional stratigraphy, eustasy and thermal maturation**

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Upper Ordovician-Upper Silurian conodont biostratigraphy, Devon Island and southern Ellesmere Island, Canadian Arctic Islands, with implications for regional stratigraphy, eustasy and thermal maturation

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Abstract: The conodont biostratigraphy for the Upper Ordovician-Upper Silurian carbonate shelf (Irene Bay and Allen Bay formations) and interfingering basinal (Cape Phillips Formation) facies is established for parts of Devon and Ellesmere Islands, central Canadian Arctic Islands. Revisions to the interpreted regional stratigraphic relationships and correlations are based on the stratigraphic distribution of the 51 conodont species representing 32 genera, identified from over 5 000 well-preserved conodonts recovered from 101 productive samples in nine stratigraphic sections. The six zones recognized are, in ascending order: *Amorphognathus ordovicianus* Local-Range Zone, *Aspelundia fluegeli* Interval Zone, *Pterospirifer celloni* Local-Range Zone, *Pt. pennatus procerus* Local-Range Zone, *Kockelella patula* Local-Range Zone and *K. v. variabilis-Ozarkodina confluens* Concurrent-Range Zone. These provided a more precise dating of the members and formations and, in particular, the range of hiatuses within this stratigraphic succession. The pattern of regional stratigraphy, facies changes, and hiatuses is interpreted as primarily related to the effects of glacio-eustasy associated with the terminal Ordovician glaciation and smaller Early Silurian glacial phases, the back-stepping of the Silurian shelf margin, and the geodynamic effects of the collision with Laurentia by Baltica to the east and Pearya to the north. Conodont Colour Alteration Index values (CAI 1–6.5) from the nine sections complement earlier graptolite reflectance data in providing regional thermal maturation data of value in hydrocarbon exploration assessments.

Keywords: Upper Ordovician-Upper Silurian, conodont biostratigraphy, Canadian Arctic

54 Islands, Allen Bay and Cape Phillips formations, thermal maturation

55 **Résumé:**

56

Draft

Introduction

The study areas lie in 1) the Vendom Fiord and Irene Bay areas, Ellesmere Island within the Central Ellesmere Fold Belt, and 2) Devon Island within both the Central Ellesmere Fold Belt and the Boothia Uplift (Fig. 1). Along the Central Ellesmere Fold Belt, the Lower Paleozoic sequence outcrops extensively and exposes a marked facies change between the carbonate shelf (Irene Bay and Allen Bay formations) and the offshore basin (Cape Phillips Formation) in the Upper Ordovician and Silurian succession. Periodically through this time interval the basinal facies partially transgressed eastward over the shelf facies. This facies relationship is of great interest for hydrocarbon exploration as massive bioherms and porous carbonate intervals, considered to be excellent reservoir rocks, are present in the shelf facies that interfinger laterally with the graptolitic shales, which are regarded as excellent source beds. The porous carbonates also host important lead-zinc deposits such as those mined earlier by Cominco (Polaris Mine) on Little Cornwallis Island. Whereas these areas have attracted various studies since the 1950s, detailed biostratigraphic work has been neglected and most publications have focused on the regional stratigraphy.

A few conodont publications have considered this stratigraphic interval in the Arctic Islands (e.g. Weyant 1968; Barnes 1974; Barnes et al. 1976; Mirza 1976; Mayr et al. 1978; Uyeno 1980, 1990; Landing and Barnes 1981; Melchin et al. 1991; Jowett 2000; Zhang et al. 2006). Among these studies, Uyeno (1990) provided relatively detailed conodont biostratigraphy, which mostly addressed the regional stratigraphy. Mirza (1976) in an unpublished M. Sc. thesis documented Late Ordovician and Silurian conodonts; the present authors are updating the taxonomic nomenclature, biostratigraphy, and revising the correlations and conclusions.

The remoteness and high cost of field operations have discouraged more detailed geological studies in these areas. In particular, there is a need for improved stratigraphic correlations to resolve: 1) the precise age of the Allen Bay Formation; 2) the chronostratigraphic relationship between the Allen Bay and Cape Phillips formations; 3) the timing of transgressive and regressive events during the Late Ordovician and Silurian; and 4) to what extent the latter are related to global eustatic changes or to tectonic events from the collisional interactions of northern Laurentia with the offshore Pearya Terrane (Hadlari et al. 2013) and Baltica (Gee et al. 2015).

This new study 1) re-examines and re-illustrates the entire conodont fauna of over 5 000 specimens from 101 productive samples from nine stratigraphic sections (Figs. 2–4; see Tables S1–S9 for section descriptions) of the Upper Ordovician to Upper Silurian succession in the Vendom Fiord area, Ellesmere Island and the Grinnell Peninsula, Devon Island; 2) identifies a total of 51 conodont species, with three in open nomenclature, belonging to 32 genera, most of which are multielement apparatuses (Figs. 5–8; see Tables S10–S16 for numerical conodont distribution data); 3) establishes the Upper Ordovician to Upper Silurian conodont biostratigraphy; 4) clarifies the age of Allen Bay Formation and that part of the Cape Phillips Formation interfingering with the Allen Bay; 5) interprets the sea level events during Late Ordovician to Late Silurian; and 6) documents the conodont Colour Alteration Index (CAI) for the faunas and the implications for the thermal maturity in the region.

Upper Ordovician and Silurian stratigraphy and sections

This study involves the Upper Ordovician to Upper Silurian succession in the Vendom Fiord area, Ellesmere Island and the Grinnell Peninsula, Devon Island, the Upper Ordovician

Irene Bay Formation and the Upper Ordovician–Upper Silurian Allen Bay Formation representing the carbonate shelf, and an interfingering Silurian unit of the basinal Cape Phillips Formation (Fig. 9)

Irene Bay Formation

Thorsteinsson (1958) established the Cornwallis Formation including basal gypsum-anhydrite, middle limestone and upper limestone-shale units. It was later raised to group status with the three units elevated to formation status namely the Bay Fiord, Thumb Mountain and Irene Bay formations (Kerr 1967). The Irene Bay Formation consists of about 83 m of recessive, greenish weathering, argillaceous limestone and minor shale. A prolific shelly fauna, informally called the “Arctic Ordovician fauna”, occurs in the Irene Bay Formation and was regarded as late Caradoc in age (Kerr 1967). This formation is the oldest stratigraphic unit dealt with by this study, occurring at sections B, 1, and 2 (Figs. 2 and 3) near the Vendom Fiord, Ellesmere Island, and sections 5, 10, 13 and 14 (Fig. 4) on Grinnell Peninsula, Devon Island. It provides an excellent marker horizon given its distinctive green weathering colour and recessive nature.

Allen Bay Formation

The Allen Bay Formation, mainly dolostone, was named and tentatively assigned an Early Silurian age by Thorsteinsson and Fortier (1954) who indicated that the formation may include Upper Ordovician strata. It was described in more detail by Thorsteinsson (1958) who designated a type section near Resolute Bay, Cornwallis Island, and correlated it to an Ashgill (Late Ordovician) to lower Wenlock (Early Silurian) interval.

The Cape Storm Formation, established by Kerr (1975), is a limestone and dolostone unit that had been included with the underlying Allen Bay Formation or with an overlying formation – either the Read Bay or the Douro. The type section is 13 km east of Cape Storm, southern Ellesmere Island, where the formation is 197 m thick. The formation was originally assigned an age of late Llandovery to early Ludlow (Kerr 1975). At its type section, it contains two members: the lower member is cliff-forming limestone, partly dolomitized, and the upper member is thin-bedded dolostone and silty dolostone, grading upward to interbedded dolostone and limestone. Thorsteinsson (1980) reported that the contact between the Allen Bay and the Cape Storm formations is situated stratigraphically a few tens of metres above an interfingering unit of the Cape Phillips Formation that yielded the graptolite *Monograptus nilssoni* (Barrande), the index species of the lowermost Ludlow graptolite zone. Therefore, the Allen Bay-Cape Storm contact was assigned to the lower Ludlow and the Cape Storm Formation was correlated to the lower-upper Ludlow.

Thorsteinsson and Mayr (1987) noted that future studies of the Cape Storm Formation on Ellesmere Island may favour excluding Kerr's lower member of the formation and including it in the underlying Allen Bay Formation. Since then, most studies (e.g. Mayr et al. 1998; de Freitas et al. 1999) have included the lower part of Cape Storm Formation in the upper part of Allen Bay Formation, correlated the Cape Storm Formation only to the lower Ludlow, and divided the Allen Bay Formation into Lower, Middle and Upper members. Mayr et al. (1998) provided detailed descriptions for the three members of the formation.

Mirza (1976) described the Late Ordovician and Silurian conodonts from the Allen Bay and Cape Storm formations. Following Thorsteinsson and Mayr's (1987) definition of the Allen Bay and Cape Storm formations, Mirza's (1976) Allen Bay and Cape Storm formations are

herein reclassified as the Lower Member of the Allen Bay Formation, and the Middle and Upper members of the Allen Bay Formation, respectively.

Section B near Vendom Fiord, southern Ellesmere Island, is the only section that exposes an almost complete Allen Bay Formation in the studied area (Fig. 2); sections 1 and 2 on southern Ellesmere Island (Fig. 3), and sections 5 and 13 on Grinnell Peninsula, Devon Island (Fig. 4) only expose the Lower Member of the formation. The Allen Bay Formation conformably overlies the Irene Bay Formation.

At section B (Fig. 2), the lower and upper parts of the Lower Member, Allen Bay Formation are composed of limestone and dolostone, respectively, with a total thickness of 357 m. The Middle and Upper members of the formation are separated by a 35 m thick interfingering unit of dark grey and black shale of the Cape Phillips Formation. These members are 301 m and 279 m in thickness, respectively, and each consists of a lower reefal facies limestone and an upper transitional facies limestone.

Cape Phillips Formation

The Cape Phillips Formation was introduced by Thorsteinsson (1958) for a sequence of dark grey to black shale, calcareous shale and minor argillaceous limestone, representing a graptolitic basin facies, with its type section located at Cape Phillips, northeastern Cornwallis Island. It was estimated to be about 3 000 m thick (Thorsteinsson and Kerr 1968) and was divided into three members (Thorsteinsson 1958). The lower, Member A, comprises mainly dolostone, argillaceous limestone, fetid shale, and cherty argillaceous limestone. The middle, Member B, conformably overlies Member A and is composed mainly of cherty argillaceous limestone, argillaceous limestone, cherty calcareous shale, and calcareous shale. The upper,

Member C, consists of an extremely monotonous succession of alternating calcareous shale, argillaceous limestone, limestone and shale. Member C accounts for roughly three-quarters of the formation's total thickness. Based on graptolite biostratigraphy, the formation was assigned a Middle Ordovician to Late Silurian age (Thorsteinsson 1958), and later modified to Late Ordovician (Ashgill) to Early Devonian (Gedinnian) (Kerr 1976; Mayr et al. 1998). More precise correlations were made by Melchin (1989), in which Members A, B, and C ranged from Late Ordovician to middle Llandovery, early to latest Telychian, and latest Telychian to Ludlow, respectively.

This present study only deals with the part of the Cape Phillips Formation that interfingers with the Allen Bay Formation at sections B (Fig. 2), 2 and 3 (Fig. 3) at Vendom Fiord, southern Ellesmere Island, and at sections 12 and 14 (Fig. 4) on Grinnell Peninsula, Devon Island.

Conodont biostratigraphy

Besides long-ranging species of *Panderodus* Ethington and *Drepanoistodus* Lindström, the Late Ordovician conodont faunas on southern Ellesmere and Devon islands are dominated by species of *Amorphognathus* Branson and Mehl that is a representative of the North Atlantic Province (Bergström 1971) with less abundant species of *Belodina* Ethington, *Pseudobelodina* Sweet and others of the North American Midcontinent Province (Sweet and Bergström 1984; Barnes et al. 1973; Barnes and Fåhraeus 1975). The Silurian conodonts tended to more cosmopolitan, and in the studied area the common Early Silurian species include those belonging to *Aspelundia* Savage, *Kockelella* Walliser, *Ozarkodina* Branson and Mehl and *Pterospirifer*

Walliser. Based on these conodonts, the following conodont zones (Figs. 2–4 and 9) are recognized.

***Amorphognathus ordovicicus* Local Range Zone**

The *Amorphognathus ordovicicus* Zone (Bergström 1971) occurs between the *Am. superbis* Zone and the Ordovician-Silurian boundary, representing almost the entire Late Ordovician Richmondian and Gamachian stages (Webby et al. 2004). *Am. ordovicicus* Branson and Mehl (Figs. 5.36–5.39) occurs in both North Atlantic and Midcontinent provinces in the Late Ordovician; hence its first appearance in the lower, but not lowermost, Richmondian Stage is a key level for global correlation (Bergström and MacKenzie 2005; Bergström et al. 2009; Bergström et al. 2011; Ferretti et al. 2014).

The existence of *Am. ordovicicus* confirms the presence of the *Am. ordovicicus* Zone in the studied area, and is supported by other relatively age-diagnostic species from the same interval such as *Culumbodina occidentalis* Sweet (Fig. 5.31), *Plegagnathus dartoni* (Stone and Furnish) (Fig. 5.20) and *Pl. nelsoni* Ethington and Furnish (Fig. 5.21). However, it needs to be discussed if this occurrence represents the entire zone interval.

Within the studied stratigraphic interval, the lowest occurrence of *Am. ordovicicus* is at the base of Irene Bay Formation at section B (Fig. 2), Vendom area, southern Ellesmere Island and at section 14 (Fig. 4), Grinnell Peninsula, Devon Island. However, this does not represent the lowest appearance of the species in the region, as this species was recovered from the upper few metres of the Thumb Mountain Formation that conformably underlies the Irene Bay Formation (Nowlan 1976). Therefore, the lowest occurrence of *Am. ordovicicus* in the Irene Bay Formation in the studied area probably occurs just above the lower boundary of *Am. ordovicicus* Zone. *Am.*

215 *ordovicicus* occurs throughout the entire Irene Bay Formation and the lower part of Lower
216 Member of Allen Bay Formation that is dominated by limestone interbedded with argillaceous
217 limestone and shale. This species disappears in the upper part of Lower Member, Allen Bay
218 Formation that is dominated by breccia dolostone. In effect, the distribution of *Am. ordovicicus*
219 tends to show that it preferred basin and perhaps more anoxic outer shelf environments; therefore,
220 its disappearance in the breccia dolomite unit in the upper part of Lower Member, Allen Bay
221 Formation is most likely due to the shallowing-upward facies change.

222 No samples collected from the Thumb Mountain Formation in this study and given the
223 facies change in the upper part of Lower Member, Allen Bay Formation, the *Am. ordovicicus*
224 Local-Range Zone only indicates its presence without clearly determining the lower and upper
225 boundaries.

227 ***Aspelundia fluegeli* Interval Zone**

228 The conodont biozonation of the Llandovery, Lower Silurian, has been constructed in
229 exceptional detail for the Telychian by Männik (1998, 2007) based on the rapid diversification of
230 species of *Pterospirifer*; however, the Rhuddanian and Aeronian biozonations remain much
231 less refined.

232 The pre-*Pterospirifer celloni* Zone was subdivided into a lower *Aspelundia expansa*
233 Zone and an upper *As. fluegeli* Zone based on the conodonts from slope and outer shelf biofacies
234 in North Greenland, and these two zones were correlated to the Rhuddanian and Aeronian,
235 respectively (Armstrong 1990). More recently, there has been a tendency to replace the *As.*
236 *fluegeli* Zone by the *Pranognathus tenuis* Zone (Cramer et al. 2011; Melchin et al. 2012); these
237 two zones are not at the exact stratigraphic level, but are roughly correlated to the graptolite *L.*

convolutus Zone in Cramer et al. (2011), or to the graptolite *pectinatus-triangulatus* Zone in Melchin et al. (2012) within Stage slice Ae2 (Fig. 9).

Given the absence of *Pr. tenuis*, *As. fluegeli* (Figs. 6.16–6.21) is used herein in determining the age of the lithostratigraphic units, with the *As. fluegeli* Interval Zone being defined by the lowest occurrence of the zonal species and the lowest occurrence of *Pterospathodus celloni* Walliser (Figs. 7.22–7.31) marking the lower and upper boundaries.

The lowest occurrence of *As. fluegeli* is at the base of the Middle Member, Allen Bay Formation, at section B (Fig. 2) and near the base of the Cape Phillips Formation at section 2 (Fig. 3), Vandom Fiord area, southern Ellesmere Island. *As. fluegeli* is a relatively long-ranging species in the studied area, occurring in almost all samples from the Middle Member, Allen Bay Formation at section B (Fig. 2), to the Cape Phillips Formation at section 2 (Fig. 3), and to a higher interval of the formation at section 14 (Fig. 4). However, the *As. fluegeli* Interval Zone is only recognized in the lower part of the Middle Member, Allen Bay Formation at section B (Fig. 2) and the lower part of the Cape Phillips Formation at section 2 (Fig. 3). Its lower boundary is marked by the lowest occurrence of the species near the base of the Middle Member, Allen Bay Formation at section B (Fig. 2) and near the base of the Cape Phillips Formation at section 2 (Fig. 3). For practical purposes, it is placed at the boundary between Lower and Middle members of the Allen Bay Formation, and between the Lower Member of Allen Bay Formation and the Cape Phillips Member at these two sections in the Vandom Fiord area, southern Ellesmere Island (Figs. 2 and 3). The *As. fluegeli* Interval Zone is not recognized on Grinnell Peninsula, Devon Island.

On Cornwallis Island (Jowett 2000), the lowest occurrence of *As. fluegeli* is within the *crispus* graptolite zone; the *As. fluegeli* Zone only covers a narrow interval of the upper *crispus* and lower *griestoniensis* graptolite zones of the Telychian (Te2). The base of the *As. fluegeli*

Interval Zone identified by this present study is temporally correlated to that of *Pranognathus tenuis* Zone (Melchin et al. 2012), and the zone covers a stratigraphic interval of middle Aeronian (Ae2) through middle Telychian (Te2) (Fig. 9), which not only covers the *Pr. tenuis* Zone, but also the overlying *Distomodus staurognathoides* and *Pt. eopennatus* zones.

The *Pt. eopennatus* Zone was established by Männik (1998) based on the collections from Estonia and Gotland, Sweden; it was later elevated to a superzone (Männik 2007). The superzone is divided into the *Pt. eopennatus* ssp. n. 1 and *Pt. eopennatus* ssp. n. 2 zones below the *Pt. celloni* Superzone. *Pt. eopennatus* Männik (Figs. 7.32–7.33) is not independently found below the *Pt. celloni* Local-Range Zone, but it co-occurs with *Pt. celloni* at section B (Fig. 2), and sections 2 and 3 (Fig. 3), which is most likely represented by morphs 3 or 2 of the Pa element; therefore, the *Pt. eopennatus* Zone is not recognized in this study. However, the *Pt. eopennatus* Superzone might occur in the upper part of the *As. fluegeli* Interval Zone. This part may be represented by an un-sampled interval between samples 319 and 367 at section B (Fig. 2), a covered interval between samples 145 and 144 at section 2 (Fig. 3).

***Pterospathodus celloni* Local-Range Zone**

The *Pterospathodus celloni* Zone was established by Walliser (1964) from the Cellon section, Carnic Alps and since recognized almost worldwide. Some attempts were made at subdividing it (e.g. Bischoff 1986; Brazauskas 1987). Notably, Männik (2007) elevated the *Pt. celloni* Zone to a superzone and divided it into three zones, i.e. *Pt. amorphognathoides angulatus*, *Pt. a. lennarti* and *Pt. a. lithuanicus* zones, which have been accepted by most recent studies involving Silurian conodont biostratigraphy (e.g. Cramer et al. 2011; Melchin et al. 2012), and correlated to the Telychian Stage slice Te3 (Cramer et al. 2011). However, in the studied area,

these three zonal species were not present whereas *Pt. celloni* (Fig. 7.22–7.31) was recovered from many samples in the Middle Member of Allen Bay Formation and the Cape Phillips Formation, Vendom Fiord area.

The interval with the total range of *Pt. celloni* is recognized as a Local-Range Zone in the study area based on the lowest and highest occurrences of the zonal species in samples 367 and 577 at section B in the Middle Member, Allen Bay Formation (Fig. 2); the *Pt. celloni* Local-Range Zone is correlated to the *Pt. celloni* Superzone (Männik 2007) (Fig. 9). Since the Cape Phillips Formation was not completely measured in the studied area, probably only the lower part of this zone occurs in the measured part of the Cape Phillips Formation at sections 2 and 3 (Fig. 3), Vendom Fiord area; it was not recognized on Devon Island.

Based on Männik (1998, 2007), the rare specimens of morphs 2 and 3 of *Pt. eopennatus* Pa element are found together with *Pt. celloni* in the lower *Pt. celloni* Superzone, which is also seen in the *Pt. celloni* Local-Range Zone at section B (Fig. 2), and sections 2 and 3 (Fig. 3) in Vendom Fiord area.

***Pterospathodus pennatus procerus* Local-Range Zone**

The *Pterospathodus pennatus procerus* Superzone was established by Jeppsson (1997) and divided into the Lower and Upper *Pt. pennatus procerus* zones based on the coniform elements. Within a wider concept, the *Pt. pennatus procerus* Superzone is useful when this division cannot be recognized (Jeppsson 1997). This has been accepted by most recent studies that have correlated the superzone to Stage slice Sh1 of the Sheinwoodian (e.g. Cramer et al. 2011; Melchin et al. 2012). Jeppsson (1997) defined the lower and upper boundaries of the *Pt. pennatus procerus* Superzone by the last appearances of *Pt. a. amorphognathoides* Walliser and

307 *Pt. pennatus procerus* (Walliser) (Figs. 7.34–7.38), respectively; therefore, it is actually an
308 interval zone.

309 Given the absence of *Pt. a. amorphognathoides* in all the measured sections, the *Pt.*
310 *pennatus procerus* Superzone is not recognized in the study. Therefore, the *Pt. pennatus*
311 *procerus* Local-Rang Zone is defined in the Cape Phillips Formation at sections 12 and 14 (Fig.
312 4), Grinnell Peninsula, Devon Island by the lowest and the highest occurrence of *Pt. pennatus*
313 *procerus* in samples 469 and 489 at section 12 (Fig. 4), respectively. However, these samples
314 probably do not represent the full local range of the species because the Cape Phillips Formation
315 was not completely measured in the study area. Therefore, this local-range zone only indicates its
316 presence without clearly established lower and upper boundaries.

317 Although *Pt. a. amorphognathoides* was not recovered from the studied sections, the
318 lower part of the defined *Pt. pennatus procerus* Local-Range Zone may be correlated to part of
319 the *Pt. a. amorphognathoides* Zone. The reasons being: 1) an interval between samples 469 and
320 479, the lower part of the measured Cape Phillips Formation at section 12 (Fig. 4), where *As. cf.*
321 *As. borenorensis* (Bischoff) (Figs. 6.22–6.28) co-occurs with *Pt. pennatus procerus*; and 2) in the
322 Cape Phillips Formation interfingering with the Irene Bay Formation and Middle Member, Allen
323 Bay Formation at section 14 (Fig. 4), where *Pt. pennatus procerus* was only recovered from
324 sample 466, but with *As. fluegeli* occurring in that sample and the samples below (468) and
325 above (465). This correlation is based on 1) the disappearance of *As. fluegeli* ssp. n. that was
326 taken as the upper boundary of the lower *Pt. a. amorphognathoides* Subzone (Männik 2007); 2)
327 the distribution of *Pt. a. amorphognathoides* and *Pt. pennatus procerus* overlaps in the upper *Pt.*
328 *a. amorphognathoides* Zone at different locations (Savage 1985; Männik 1998; Jowett 2000), or
329 almost overlaps within the *Pt. a. amorphognathoides* Zone (Walliser, 1964; Corradini et al.

2015); and 3) the juvenile specimens of *Pt. a. amorphognathoides* and *Pt. pennatus procerus* are similar to each other, and the juvenile specimens of *Pt. pennatus procerus* (Fig. 7.36) identified by this study perhaps could be assigned to *Pt. a. amorphognathoides*.

It is worth noting that samples 577 and 601 in the upper part of Middle Member, Allen Bay Formation at section B (Fig. 2) contain *Ps. bicornis* Drygant (Fig. 8.7), and both *Pt. celloni* and *Ps. bicornis* co-occur in the same sample (577). This co-occurrence has not been reported elsewhere. Globally, *Pt. celloni* does not extend into the *Pt. a. amorphognathoides* Zone, but the lowest occurrence of *Ps. bicornis* can be found in the lower *Pt. a. amorphognathoides* Zone (Jeppsson 1997; Corradini 2007; Männik 2007). Therefore, the co-occurrence of the two species in the study area would suggest that the “*Ps. bicornis*” interval at section B is close to the boundary between the *Pt. celloni* and *Pt. a. amorphognathoides* zones. Since the lower part of *Pt. pennatus procerus* Local-Range Zone is correlated to the *Pt. a. amorphognathoides* Zone as discussed above, this “*Ps. bicornis*” interval at section B is questionably correlated to the lower *Pt. pennatus procerus* Local-Range Zone (Fig. 2).

***Kockelella patula* Local-Range Zone**

The *Kockelella patula* Zone was established by Walliser (1964) at the Cellon section, Austria where it either directly succeeds the *Pt. amorphognathoides* Zone (Walliser 1964), or lies within a gap recognized between the two zones (Corradini et al. 2015). Whereas *K. patula* Walliser dominated that Cellon fauna (Walliser 1964; Corradini et al. 2015), it has not been found in most studied sequences worldwide. A detailed study of latest Telychian, Sheinwoodian and early Homerian conodonts by Jeppsson (1997) identified the *Kockelella ranuliformis*, *Ozarkodian sagitta rhenana*, and lower and middle *K. walliseri* zones (Fig. 9) between the *Pt.*

pennatus procerus and *K. patula* zones. Given the rare occurrence of *K. patula*, the *K. patula* Zone tends to have been abandoned in recent studies (e.g. Cramer et al. 2011; Melchin et al. 2012). Based on Jeppsson (1997) and Cramer et al. (2011), the *K. patula* Zone can be correlated to upper *K. walliseri* Zone and Stage slice lower Sh3 of the Sheinwoodian.

K. patula (Fig. 7.19–7.21) was only recovered from the Cape Phillips Formation in the upper part of section 12 (Fig. 4), Grinnell Peninsula, Devon Island. The *K. patula* Local-Range Zone is based on the lowest and highest occurrence of the zonal species in samples 489 and 497 (Fig. 4). Herein, it is questionably correlated to the *K. ranuliformis*, *Ozarkodina sagitta rhenana*, and *K. walliseri* zones (Cramer et al. 2011) that occur above the *Pt. pennatus procerus* Local-Range Zone and to the Stage slice from uppermost Sh1 to lower Sh3 of the Sheinwoodian (Fig. 9), for the following reasons: 1) the world-wide total range of *K. patula* is poorly known, owing to its rare occurrence; 2) the lowest occurrence of *K. patula*, although lacking Pa element, and the highest occurrence of *Pt. pennatus procerus* co-occur in the same sample (489) at section 12 (Fig 4), which makes the lowest occurrence of the zonal species questionable; and 3) sample 489, barren sample 490, and a covered interval above 490 may be related to the *K. ranuliformis*, *Ozarkodina sagitta rhenana*, and lower and middle *K. walliseri* zones (Jeppsson 1997).

***Kockelella v. variabilis*-*Ozarkodina confluens* Concurrent-Range Zone**

The *Kockelella v. variabilis* Interval Zone, as used by Cramer et al. (2011) and Melchin et al. (2012), occurs above the *K. crassa* and below the *Ancoradella ploeckensis* zones, and is correlated to Stage slice upper Go1 and Go2 of the Gorstian (Fig. 9).

K. v. variabilis Walliser (Fig. 7.8) was only recovered from two samples (671 and 775) in the lower part, representing the reefal facies, of the Upper Member, Allen Bay Formation at

section B (Fig. 2), Vandom Fiord area, southern Ellesmere Island, which supports the presence of the *K. v. variabilis* Interval Zone in the studied area. However, the total stratigraphic distribution of *K. v. variabilis* is not only restricted to the *K. v. variabilis* Interval Zone, but ranges from the base of the *K. crassa* Zone to the *Pedavis latialata* Zone (roughly equal to the *Ozarkodina snajdri* Interval Zone in Fig. 9) based on Sweet (1988). Within this interval, *K. v. variabilis* co-occurs with *Ozarkodina confluens* (Branson and Mehl) (Fig. 6.29) (Sweet 1988), which is also present in section B (Fig. 2). Neither *K. crassa* (Walliser) nor *Ancoradella ploeckensis* Walliser was found in the studied area; therefore, it is uncertain if the total range of *K. v. variabilis* at section B is restricted only to the *K. v. variabilis* Interval Zone. Given the co-occurrence of *K. variabilis* and *O. confluens*, this study establishes the *K. v. variabilis*-*O. confluens* Concurrent-Range Zone and correlates it to both the *K. crassa* Zone and *K. v. variabilis* Interval Zone, and to the entire Gorstian (Fig. 9).

Age of the three members of the Allen Bay Formation and the interfingering unit of the Cape Phillips Formation

The upper boundary of Allen Bay Formation was placed in the lower Ludlow, Upper Silurian by Thorsteinsson (1980 with contributions by Uyeno), based on graptolites and conodonts, and the lower boundary of the formation was assigned to the upper Richmondian, Upper Ordovician by Uyeno (1990), based on conodonts. These correlations have been followed by later studies (e.g. Mayr et al. 1998; de Freitas et al. 1999). The three members of the Allen Bay Formation and the disconformities between them were identified by all these studies; however, the ages of these members and the extent of the stratigraphic gaps that the disconformities represent have not been well documented.

Lower Member of the Allen Bay Formation

The Lower Member of the Allen Bay Formation contains the *Amorphognathus ordovicicus* Local-Range Zone that probably ranges into lower Richmondian, Upper Ordovician (Fig. 9), but not the lowest, because the zonal species also occurs in the underlying uppermost Thumb Mountain and Irene Bay formations. It is uncertain whether the age of this member ranges higher into the late Richmondian and Gamachian.

At section B (Fig. 2), *Am. ordovicicus* together with *Belodina confluens* Sweet (Figs. 5.7–5.9) occurs in the lower part of Lower Member; however, the latter species continues into the middle part of the Lower Member where the former disappears.

Generally in the North American Midcontinent Province, *Belodina confluens* (zonal species of the *B. confluens* Zone) ranges from Edenian to lower Richmondian, and only co-occurs with *Am. ordovicicus* in a short interval within the *Oulodus robustus* Zone, or the lower *Am. ordovicicus* Zone (Sweet 1988). However, at section B (Fig. 2), Vendom Fiord, southern Ellesmere Island, this species not only co-occurs with *Am. ordovicicus* in the Irene Bay Formation and lower limestone unit of the Lower Member, Allen Bay Formation, but also exists in the upper breccia dolostone unit of the Lower Member, Allen Bay Formation where *Am. ordovicicus* is absent. This may be interpreted either as the longest range of *B. confluens* in North America or, more likely, as the limited stratigraphic range of *Am. ordovicicus* in the studied area. Thus, the Irene Bay Formation and the Lower Member of the Allen Bay Formation are considered to probably lie within the lower *Am. ordovicicus* Zone recognized by GTS (2012) (Fig. 9).

The genus *Gamachignathus* McCracken, Nowlan and Barnes was reported from the lower part of the Allen Bay Formation in central-eastern Cornwallis Island (McCracken, pers.

comm. 1987 in Uyeno 1990), but the upper part of the Lower Member, Allen Bay Formation at most measured sections in the study area is barren of conodonts except for a few samples containing *B. confluens* and other non-zonal simple cone species at section B. Therefore, it is most likely that: 1) strata representing the upper Richmondian and Gamachian are absent in the studied area; 2) the early Richmondian is the lower age limit of the disconformity between the Lower and Middle members of the Allen Bay Formation; and 3) the major Late Ordovician regression in this region began earlier than the graptolite *fastigatus/persculptus* Zone as interpreted by de Freitas et al. (1999).

Middle Member of the Allen Bay Formation

The *Aspelundia fluegeli* Interval Zone, *Pterospathodus celloni* Local-Range Zone and possibly the lower *Pt. pennatus procerus* Local-Range Zone are recognized within the Middle Member, Allen Bay Formation, which is correlated to the Stage slice Ae2 and Ae3 of the Aeronian, and Te1 to Te5 of the Telychian. The lower boundary of the *As. fluegeli* Interval Zone and the upper boundary of the underlying *Amorphognathus ordovicicus* Local-Range Zone define a stratigraphic gap between the Lower and Middle members of the Allen Bay Formation, which probably ranges from upper Richmondian through Rhuddanian (Rh1–Rh3) to lower Aeronian (Ae1) (Fig. 9).

The conodont fauna within the *As. fluegeli* Interval Zone is not abundant; besides the zonal species, *Dapsilodus* sp. (Figs. 8.1–8.3) occurs, which is only present in the Silurian in the study area, and also a few other coniform species (mainly panderodontids) surviving the Late Ordovician mass extinction (Fig. 2). This fauna represents the pioneer community during the initiation of the Early Silurian transgression onto the platform, probably during the Aeronian Ae2,

or even during the Telychian (Te2), considering the lowest occurrence of *As. fluegeli* on Cornwallis Island (Jowett 2000), rather than Rhuddanian as interpreted by de Freitas et al. (1999).

This transgression was more extensive in the middle Telychian (Te3) as represented by the *Pt. celloni* Local-Range Zone (Fig. 9). This is shown by: 1) the conodont fauna within the *Pt. celloni* Local-Rang Zone is much more abundant and diverse than that within the underlying *As. fluegeli* Interval Zone; important species for this interval, besides the zonal species, include *Apsidognathus t. tuberculatus* Walliser (Fig. 7.13), *Ap. t. lobatus* Bischoff (Figs. 7.9–7.10), *Astropentognathus irregularis* Mostler (Figs. 7.1–7.7), *Aulacognathus angulatus* Bischoff (Fig. 7.16), *Au. bullatus* (Nicoll and Rexroad) (Figs. 7.17–7.18), and *Pt. eopennatus* (Figs. 7.32–7.33); and 2) the *Pt. celloni* Local-Range Zone is recognized in the interfingering Cape Phillips Formation unit, a basinal facies laterally equivalent with the Middle Member, Allen Bay Formation, at section 3 (Fig. 3). Therefore, the Middle Member of the Allen Bay Formation was deposited during the extensive transgressive event in the Early Silurian, with the age of this member being from Aeronian (Ae2) to late Telychian (Te4 and possible Te5).

Interfingering unit of the Cape Phillips Formation between the Middle and Upper members, Allen Bay Formation

Section B on southern Ellesmere Island contains a complete section of the Allen Bay Formation, and also includes a 35 m interval of dark gray and black shale of the Cape Phillips Formation that interfingers between the Middle and Upper members (Fig. 2). This unit represents a change from shelf to basin facies, and probably represents the maximum transgression that was initiated in the middle Aeronian. With the lack of carbonates, only one sample (644) was

collected from this Cape Phillips unit. Only *Panderodus unicostatus* (Branson and Mehl) (Figs. 8.25–8.31) and *Wurmiella e. excavata* (Branson and Mehl) (Figs. 6.37–6.41) are present. This latter species ranges from the *Pt. celloni* Local-Range Zone in the Middle Member to the *K. v. variabilis*-*O. confluens* Concurrent-Range Zone in the Upper Member, Allen Bay Formation at section B (Fig. 2), and from the *Pt. pennatus procerus* Local-Range Zone to the *K. patula* Local-Range Zone in the Cape Phillips Formation at section 12, Grinnell Peninsula, Devon Island (Fig. 4).

Because of the incomplete measurement of the Cape Phillips Formation (beyond the 35 m unit) in the studied area, several conodont zones are not recognized from upper Sheinwoodian to Homerian (Fig. 9). This does not necessarily mean that the strata formed during this time interval are not represented within the Cape Phillips Formation, since no unconformity has been recognised within the formation. Therefore, this 35 m thick shale unit of Cape Phillips between the Middle and Upper members, Allen Bay Formation at section B probably has an age of earliest Sheinwoodian (Sh1) to the end of Homerian (Ho3) when the maximum transgression caused the shelf facies to be replaced by the basin facies. This facies replacement was initiated in the earliest Sheinwoodian (Sh1), which is slightly later than a major transgression during the *Pt. amorphognathoides* Zone interval reported by de Freitas et al. (1999). The possibility of a paraconformity between the unit and the overlying Upper Member cannot be ruled out.

Upper Member of the Allen Bay Formation

As noted above, at Section B the Upper Member, Allen Bay Formation overlies the 35 m unit of the Cape Phillips Formation (Fig. 2) that extends onto the shelf during a period of maximum transgression. The *Kockelella v. variabilis*-*Ozarkodina confluens* Concurrent-Range

Zone is the only conodont zone recognized in the carbonate unit immediately above this shale unit (Figs. 2 and 9). It occurs in the lower part of the Upper Member, Allen Bay Formation and is correlated to the Gorstian (Fig. 9). The upper part of the Upper Member, Allen Bay Formation only yields *Panderodus unicostatus*, so it is uncertain whether this upper part belongs to the same or other zones of Ludfordian age. It is possible that the strata above the *K. v. variabilis-O. confluens* Concurrent-Range Zone belong to the Ludfordian or lower Ludfordian. Without strong supporting evidence, this study follows de Freitas et al. (1999) in correlating the upper boundary of the Upper Member, Allen Bay Formation to the upper boundary of the Gorstian (Fig. 9).

The carbonates of the Upper Member, Allen Bay Formation at section B represent a regression that resulted in the basin facies retreating from the shelf settings. A further major transgression in the early Ludfordian, recognized by de Freitas et al. (1999), is represented by the Cape Phillips shale on the top of the Upper Member, Allen Bay Formation (Fig. 2).

Interpreted patterns of eustasy and paleoceanography during the Early Silurian in the central Arctic Islands, with comparisons to other key regions in Canada

The details of the stratigraphy and conodont faunas reported herein permit an elaboration on the interpretations of the regional patterns of eustasy and paleoceanography for the central Arctic Islands and comparisons with other key documented areas in Canada, representative of northern Laurentia.

The main eustatic events and trends are:

a) sea level remained relative high during the early Richmondian, represented by the Irene Bay Formation and Lower Member, Allen Bay Formation;

b) a major late Ordovician regression is marked by a hiatus in the Arctic succession between the Lower and Middle members, Allen Bay Formation, partly representing the Hirnantian glaciation on northern Gondwana, but in this region extending through the Rhuddanian and early Aeronian (Ae1);

c) a modest transgression persisted through the Aeronian (Ae2) (or the Telychian (Te2)) to the late Telychian (Te5) that is reflected by the facies changes documented herein for the Middle Member, Allen Bay Formation;

d) a more significant transgression starting in the early Sheinwoodian (Sh1) is marked by the interfingering 35 m unit of Cape Phillips Formation shale assigned to an interval within the earliest Sheinwoodian (Sh1) to the end of Homerian (Ho3); and

e) a regressive phase is marked by the Upper Member, Allen Bay Formation during the Gorstian and possibly into the early Ludfordian.

These patterns do not readily match some of the interpreted broad global Silurian eustatic patterns advocated, for example, by Loydell (1998), Johnson (2006), and Haq and Schutter (2008) and compared in Trotter et al. (2016), namely: transgression during the early Rhuddanian; transgressive-regressive oscillations in the Aeronian-early Telychian; regressive phases within the Wenlock; and transgression during the early Ludlow. This region may have been affected by regional geodynamic effects resulting from the collision of Baltica with Laurentia to the east (Pollock et al. 2007; Gee et al. 2015) and the docking of Pearya to the north (Hadlari et al. 2013) to create regional differences in apparent sea level changes. These may have generated more significant regional eustatic effects than those induced by minor glacial re-advances on northern Gondwana during the Early Silurian.

The key paleoceanographic patterns and events of the area include the restricted circulation on the carbonate platform, a poorly rimmed reefal bank margin at times, and the relatively deep and anoxic offshore shale basin. Expressions of oceanographic changes include: the transgressions and regressions influenced by oceanic thermal expansion during warm phases; back-stepping of the carbonate margin allowing transgression of the basinal facies (Cape Phillips unit; de Freitas et al. 1999); and the broad geodynamic effects related to the docking of Baltica to eastern Laurentia during the Silurian and the Pearya Terrane against the northern Innuitian margin. A key question is the formation of the 35 m unit of Cape Phillips shale within the platform Allen Bay facies. The most accepted explanation is through the back-stepping of the carbonate margin with the consequent eastward migration of the basinal shale facies. It could partly be a product of the shut-down of the carbonate factory during a cooling phase in the Wenlock (e.g. Trotter et al. 2016, fig. 3). Changes in the regional oceanographic circulation with the docking of Pearya to the north could also have affected the pattern of upwelling of anoxic waters onto the carbonate platform (cf. Servais et al. 2014), perhaps accentuated near the sharp angular change in orientation of the margin (Fig. 1).

In a wider context, it is possible to draw comparisons with other areas of northern Laurentia that preserve a good, well documented, stratigraphic and conodont biostratigraphic record for the Late Ordovician-Early Silurian. The changing eustasy strongly controls the overall paleogeography of the epeiric seas in relation to areas of exposed Canadian Shield.

To the south-east of the Arctic Islands, samples from both wells and outcrops from the Hudson Bay Basin and Foxe Basin provided a stratigraphic and conodont biostratigraphic framework (Zhang and Barnes 2007; Zhang 2011, 2013). This demonstrated the presence of a regional hiatus for the late Richmondian-Gamachian to early Rhuddanian interval (Zhang and

Barnes 2007, fig. 2; Zhang 2011, fig. 1; 2013, fig. 7), starting at a similar time to the Devon/Ellesmere islands sequences but with sedimentation starting earlier in the Rhuddanian rather than the early Aeronian. Lateral facies shifts were also present during the Telychian-Wenlock (Zhang and Barnes 2007, fig.2), probably equivalent to those found in Devon/Ellesmere islands but more likely produced by glacio-eustatic processes.

Further to the south-east is the Anticosti Basin, where extensive stratigraphic and conodont studies were undertaken for the Late Ordovician to Telychian interval (e.g, Nowlan and Barnes 1981; McCracken and Barnes 1981; Uyeno and Barnes 1983; Zhang and Barnes 2002, 2004). Here, the hiatus near the Ordovician-Silurian boundary is of minor duration, lying above a thick Gamachian carbonate sequence (see also Bergström et al. 2011). The subtle eustatic changes through most of the Llandovery have been demonstrated through conodont community statistical analyses (Zhang and Barnes 2004; Zhang et al. 2006).

Far to the south-west of the Arctic Islands, the sequences occur in the northern and central Canadian Rocky Mountains. Detailed platform-to-basin transects (Pyle and Barnes 2002, 2003; Zhang et al. 2005) have demonstrated the significant hiatus from the latest Ordovician to the early Aeronian, with the Late Ordovician platform carbonates of the Beaverfoot and Robb formations being slightly older than the latest Ordovician Ospika Formation in the basinal facies to the west.

These various conodont biostratigraphic studies from other major depositional settings across thousands of kilometres of northern Laurentia, when combined with those from the central Canadian Arctic Islands, demonstrate that the eustatic lowstand associated with the peak Gamachian/Hirnantian glaciation affected the entire area. In the centre of the craton in the Hudson Bay Basin, the hiatus occupies most of the Gamachian with renewed deposition marked

by the early Rhuddanian Severn River Formation. This early Llandovery transgression has different ages, being earliest in the Anticosti Basin probably due to it being a subsiding basin. The carbonate shelves near the continental margins of the central Arctic Islands and northern and central Rocky Mountains were probably additionally influenced by regional geodynamic processes with the longer hiatus typically ranging from Gamachian through to Aeronian.

Subtle eustatic and paleoclimatic changes for the early Silurian are well documented particularly for Baltica, and have been referred to as primo and secundo episodes and events (e.g. Aldridge et al. 1993; Jeppsson 1998; Trotter et al. 2016). The limited conodont abundance and presence of hiatuses in the central Arctic Islands described here do not permit a detailed comparison with these events.

Regional thermal maturation values using the conodont Colour Alteration Index (CAI)

Of interest to exploration for hydrocarbons is the regional pattern of thermal maturation. This can be assessed from changes to the organic matter in the phosphatic hard tissue of conodonts (Epstein et al. 1977; Mayr et al. 1978; Legall et al. 1981) and also from the organic periderm of graptolites (Goodarzi et al. 1992; Gentzis et al. 1996).

The conodont species and their abundance in each sample for this present study are reported in Tables S10–S16, with the conodont Colour Alteration Index (CAI) value(s) noted at the top of each table and their regional distribution in Figure 1. CAI values range from 1–6.5, representing a significant range of burial temperatures. The lowest values (CAI 1–3) are at Sections 10, 12, 13 and 14 on Grinnell Peninsula, Devon Island as well as at Section 5 nearby on northwest Devon Island. These are all within or adjacent to the Boothia Uplift that separates the

Parry Island and Central Ellesmere fold belts and similar values are found further south on Cornwallis Island (Jowett 2000) along the axis of this positive tectonic feature. Sections 5, 12 and 14 show CAI values of 1–2 and Sections 10 and 13 exhibit CAI values of 2–3 (Fig. 10; Tables S14–S16) with the latter possibly affected more by local faulting. These represent burial temperatures in the range of 50°C–140°C (CAI 1–2) and 60°C–200°C (CAI 2–3), respectively. To the north-east, 200–500 km along the Central Ellesmere Fold Belt at Sections 2 and 3 (Hoved Island, and where Mayr et al. (1978) initially reported maturation data for nearby Bjorne Peninsula) and at Section 1 (north-east of Irene Bay) the CAI values increase to 3–4 (110°C–300°C). These reflect the greater level of tectonic deformation and perhaps burial depth. The highest CAI values of 5–6 (300°C–550°C), locally even 6.5 (440°C–610°C), are at Section B at Vendom Fiord, with two small parts of the section having lower values of 4–5 (Fig. 10; Tables S10–S13). Vendom Fiord, 20 km east of Hoved Island, marks the axis of tightly folded strata and close to the Jones Sound Fold Belt and the Inglefield (Bache) Uplift that occur along much of the east coasts of Devon and Ellesmere Island (Fig. 1). Similar CAI values of 5 were reported in Trettin (1994) for the Lower Paleozoic rocks in northern Ellesmere Island.

Some studies of Arctic graptolites have reported on inferred burial temperatures and maturation. Mean maximum graptolite reflectance values from numerous sections range from 0.6% in Cornwallis Island and northwestern Devon Island to 4.7% in northern and central Ellesmere Island (Gentzis et al. 1996). This lateral reflectance variation was attributed to differing burial depths and tectonic loading of the graptolite-bearing strata primarily beneath a thick Devonian synorogenic siliciclastic cover.

This significant thickness of Devonian clastics that was shed over this region from the east was related to the final closure of Baltica with Laurentia, generating the East Greenland

Caledonides and the Acadian Orogeny (Trettin et al. 1991; Trettin 1994; Mayr et al. 1998; Gentzis et al. 1996; Gee et al. 2015). About 4–7 km of Late Silurian-Carboniferous deposits accumulated in this studied area, with about 3 km since removed by erosion; however, only about 2 km of strata accumulated in the Boothia Uplift area. An estimated 12 km of Mesozoic and Cenozoic evaporites and clastics filled the adjacent Sverdrup Basin to the west (Fig. 1), but most of that thickness did not extend to the eastern margin of the basin and had little effect in the study area. A mild orogenic phase occurred with the Cornwallis Disturbance that elevated the Boothia Uplift, followed by the Ellesmerian Orogeny (latest Devonian–earliest Carboniferous), and later rifting that established the Sverdrup Basin, which was deformed by the Eurekan Orogeny (Eocene-Oligocene) (Trettin 1991; Mayr et al. 1998).

Thus, the thermal maturation patterns described herein (Fig. 1) are likely to have been produced mainly by the regional variations in tectonic stacking during phases of deformation and particularly through burial by the foreland clastic wedge created by the Ellesmerian Orogeny, with some areas receiving only minor maturation levels given the buttressed protection of the Boothia Uplift. In summary, these conodont CAI data document areas exhibiting values of CAI 1–3 (Fig. 1) that lie within the wet gas to oil window that could be prospective for hydrocarbon exploration. Areas where CAI values are 4–6.5 (Fig. 1) are mainly above dry gas generation and are not prospective for such exploration.

Summary

The Lower Paleozoic stratigraphic succession for the Innuitian Orogen is best exposed on Devon and Ellesmere Islands, central Canadian Arctic Islands. The carbonate shelf facies passes westwards at the ancient shelf margin into the basinal shale facies. Later tectonic phases resulted

in some areas having limited deformation (Boothia Uplift) and others with strong folding (Parry Island and Central Ellesmere fold belts). These geological complexities, combined with the area being remote and expensive for field logistics, have resulted in mostly reconnaissance studies with limited specialized research investigations.

Special logistic opportunities allowed this study of key stratigraphic sections with the collection of samples for conodont biostratigraphy. Over 5 000 conodont specimens were recovered from 101 productive conodont samples and taxonomic study identified 51 species representing 32 genera, with three in open nomenclature. Based on the faunas the key zones recognized are, in ascending order: *Amorphognathus ordovicicus* Local-Range Zone, *Aspelundia fluegeli* Interval Zone, *Pterospathodus celloni*, *Pt. pennatus procerus* and *Kockelella patula* Local-Range zones, and *Kockelella v. variabilis*-*Ozarkodina confluens* Concurrent-Range Zone.

The conodont biostratigraphic data establish the ages of the main stratigraphic units as: 1) Irene Bay Formation and Lower Member, Allen Bay Formation – early Richmondian, Late Ordovician; 2) Middle Member, Allen Bay Formation - Aeronian (Ae2) to late Telychian (Te5), Llandovery, Early Silurian; 3) interfingering unit of Cape Phillips Formation - early Sheinwoodian (Sh1) to late Homerian (Ho3), Wenlock, Early Silurian; and 4) Upper Member, Allen Bay Formation - Gorstian, possibly extending into the early Ludfordian, Late Silurian. Major hiatuses occur above the Lower Member, Allen Bay Formation and possibly above the interfingering Cape Phillips unit.

Five main eustatic events and trends are recognized: a) a relatively high sea level represented by the Irene Bay and Lower Member, Allen Bay Formation (early Richmondian); b) a major late Ordovician-early Silurian regression marked by a hiatus between the Lower and Middle members, Allen Bay Formation (Hirnatian to early Aeronian); c) a modest transgression

(Aeronian (Ae2) to late Telychian (Te4/Te5)) marked by the Middle Member, Allen Bay Formation; d) a more significant transgression (early Sheinwoodian (Sh1)), marked by the interfingering 35 m unit of Cape Phillips Formation shale (Sheinwoodian (Sh1) to the end of the Homerian (Ho3)); and e) a regressive phase marked by the Upper Member, Allen Bay Formation (Gorstian and possibly to early Ludfordian).

These patterns show some differences to the interpreted global Silurian eustatic patterns, possibly because of regional geodynamic effects resulting in apparent sea level changes from the collisions with Laurentia by Baltica to the east and Pearya to the north. Key paleoceanographic patterns and events in the area include the restricted circulation on the carbonate platform, a partly rimmed reefal bank margin at times with eastward backstepping to produce the interfingering Cape Phillips shale unit, and the relatively deep and anoxic offshore shale basin to the west.

The conodont CAI values at the nine stratigraphic sections ranging between 1 and 6.5 are compared with the thermal maturation data established by earlier graptolite reflectance studies. The conodont thermal maturation patterns are interpreted to reflect the regional variations in tectonic stacking during later phases of deformation and particularly through burial by the foreland clastic wedge created by the Ellesmerian Orogeny (late Devonian–earliest Carboniferous), but with some areas having low maturation levels as a result of the buttressed protection of the Boothia Uplift. Those areas exhibiting values of CAI 1–3 lie within the wet gas to oil window and could be prospective for hydrocarbon exploration.

Acknowledgements

This study was supported by research grants to Chris Barnes from the Natural Sciences and Engineering Council of Canada (NSERC) and the Geological Survey of Canada. Field logistic support and advice was kindly given to Chris Barnes by Panarctic Oil Company, the Geological Survey of Canada (GSC), and the Polar Continental Shelf Project. Additional stratigraphic data and samples were provided to Khusro Mirza by Sproule Associates Ltd., Calgary. Shunxin Zhang acknowledges continued support from the Strategic Investments in Northern Economic Development (SINED) and the Canada–Nunavut Geoscience Office (CNGO) for Arctic geoscience research. Thanks are extended to Pat Hunt in GSC, Ottawa and Jianqun Wang in the Carleton University who helped in taking the SEM images, to Sandy McCracken, Peep Männik, and an anonymous reviewer who acted as scientific reviewers, and to Ali Polat, Jisuo Jin, and Brenda Tryhuba who edited the manuscript.

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Figure Captions

Fig. 1. Geological map of Devon Island and southern Ellesmere Island with index map showing the different tectonic units among the Canadian Arctic Islands and the location of the studied area within the Franklinian Mobile Belt (modified from Trettin 1991). Dots with different colours represent both section localities and conodont Colour Alteration Index (CAI) values. Yellow, red and black dots represent CAI values 1–3, 3–4, and 4–6.5, respectively.

Fig. 2: Conodont distribution in the Irene Bay and Allen Bay formations at section B, southern Ellesmere Island. See Fig. 1 for location, Fig. 3 for lithologic legend, Table S1 for section description, and Tables S10 and S11 for numerical distribution data. C-R: Concurrent-Range; L. *Pt. p. p.*: Lower *Pt. pennatus procerus* Local-Range Zone; Z.: Zone; C. P.: Cape Phillips.

Fig. 3. Conodont distribution in the Irene Bay, Allen Bay and Cape Phillips formations at sections 1–3, southern Ellesmere Island. See Fig. 1 for locations, Tables S2–S4 for section descriptions and Tables S12–S14 for numerical distribution data. L-R: Local-Range.

Fig. 4. Conodont distribution in the Irene Bay, Allen Bay and Cape Phillips formations at sections 5, 10 and 12–14, Grinnell Peninsula, Devon Island. See Fig. 1 for location, Fig. 3 for lithologic legend, Tables S5–S9 for section descriptions and Tables S14–S16 for numerical distribution data. L-R: Local-Range.

Fig. 5. Ordovician conodonts (all illustrated specimens in Figs. 5–8 and 10 are curated in the National Type Collection of Invertebrate and Plant Fossils, the Geological Survey of Canada (GSC), Ottawa, Ontario; GSC##### is curation number). **1–3. *Besselodus borealis* Nowlan and McCracken** (×80); from 451, section 13; 1. lateral view of Sa element, GSC138320; 2. lateral view of Sb-c element, GSC138321; 3. lateral view of M element, GSC138322. **4–6. *Paroistodus? mutatus* (Branson and Mehl)** (×65); from 451, section 13; 4. lateral view of M

935 element, GSC138323; 5. lateral view of Sa element, GSC138324; 6. lateral view of Sb-c element,
 936 GSC138325. **7–9. *Belodina confluens* Sweet** ($\times 80$ except 9×50), from 451, section 13; 7. outer
 937 lateral view of eobelodiniiform element, GSC138326; 8. inner lateral view of compressiform
 938 element, GSC138327; 9. outer lateral view of grandiform element, GSC138328. **10–11.**
 939 ***Staufferella* n. sp. A McCracken** ($\times 50$); from 0, section B; 10. posterior view of symmetric
 940 element, GSC138329; 11. posterior view of asymmetric element, GSC138330. **12–14.**
 941 ***Panderodus breviusculus* Barnes** ($\times 50$); from 0, section B; 12, outer lateral view of graciliform
 942 element, GSC138331; 13. inner lateral view of arcuatiform element, GSC138332. 14. inner
 943 lateral view of compressiform element, GSC138333. **15–17. *Pseudobelodina? dispansa***
 944 **(Glenister)** ($\times 80$); from 451, section 13; 15. outer lateral view of Sc^1 element, GSC138334; 16.
 945 inner lateral view of Sg^2 element, GSC138335; 17. inner lateral view of Sg^1 element,
 946 GSC138336. **18–19. *Pseudobelodina v. vulgaris* Sweet** ($\times 80$); from 451, section 13; 18. inner
 947 lateral view of Sc^0 element, GSC138337; 19. inner lateral view of Sg^2 element, GSC138338. **20.**
 948 ***Plegagnathus dartoni* (Stone and Furnish)** ($\times 45$); from 160, section B; inner lateral view,
 949 GSC138339. **21. *Plegagnathus nelsoni* Ethington and Furnish** ($\times 50$); from 451, section 13;
 950 inner lateral view of nelsoniform element, GSC138340. **22. *Pseudooneotodus mitratus***
 951 **(Moskalenko)** ($\times 65$); from 451, section 13; upper view, GSC138341. **23–26. *Drepanoistodus***
 952 ***suberectus* (Branson and Mehl)** ($\times 50$); from 451, section 13; 23. lateral view of oistodiform,
 953 GSC138342; 24. lateral view of homocurvatiiform element, GSC138343; 25. lateral view of
 954 curvatiiform element, GSC138344; 26. lateral view of suberectiform element, GSC138345. **27–**
 955 **28. *Zanclodus* sp.** ($\times 80$); 27. from 130, section B; inner lateral view of long base element,
 956 GSC138346; 28. from 451, section 13; inner lateral view of short base element, GSC138347. **29–**
 957 **30. *Pseudobelodina torta* Sweet** ($\times 60$); from 0, section B; 29. inner lateral view of Sg^1 element,

958 GSC138348; 30. outer lateral view of Sc⁰ element, GSC138349. **31. *Culumbodina occidentalis***
 959 **Sweet** (×45); from 0, section B; inner lateral view of denticulate element, GSC138350. **32–34.**
 960 ***Plectodina tenuis* (Branson and Mehl)** (×55); from 0 (except 34 from 80), section B; 32.
 961 posterior view of Pb element, GSC138351; 33. inner lateral view of M element, GSC138352; 34.
 962 inner lateral view of Sc element, GSC138353. **35. *Coelocerodontus trigonius* Ethington** (×80);
 963 from 80, section B; posterior-lateral view of tetragonal element, GSC138354. **36–39.**
 964 ***Amorphognathus ordovicicus* Branson and Mehl** (×65 except 37×45); from 451, section 13
 965 (except 39 from 0, section B); 36. lateral view of S element, GSC138355; 37. upper view of Pa
 966 element, GSC138356; 38. outer lateral view of Pb element, GSC138357; 39. posterior-lateral
 967 view of M element, GSC138358.
 968 **Fig. 6.** Silurian conodonts. **1–3. *Oulodus* sp.** (×45); from 478 (except 1 from 476), section 12; 1.
 969 inner lateral view of Pb element, GSC138359; 2. inner lateral view of Sc element, GSC138360; 3.
 970 posterior view of Sb element, GSC138361. **4–6. *Rexroadus* cf. *R. kentuckyensis* (Branson and**
 971 **Branson)** (×70); from 145, section 2; 4. posterior view of Sb element, GSC138362; 5. lateral
 972 view of Pa element, GSC138363; 6. inner lateral view of Sc element, GSC138364. **7–10.**
 973 ***Oulodus confluens* (Branson and Mehl)** (×65 except 10 ×50); from 525, section B; 7. posterior
 974 view of Sa element, GSC138365; 8. inner lateral view of Sc element, GSC138366; 9. posterior
 975 view of M element, GSC138367; 10. posterior view of Sb element, GSC138368. **11–15.**
 976 ***Distomodus staurognathoides* (Walliser)** (×55 except 12 ×75; 15 ×35); from 130b, section 3; 11.
 977 inner lateral view of Pb element, GSC138369; 12. posterior-lateral view of Sa element,
 978 GSC138370; 13. outer lateral view of Sc element, GSC138371; 14. upper view of Pa element,
 979 GSC138372; 15. posterior-lateral view of Sb element, GSC138373. **16–21. *Aspelundia fluegeli***
 980 **(Walliser)** (×60); from 129, section 3; 16. inner lateral view of Pb element, GSC138374; 17.

981 inner lateral view of Sc element GSC138375; 18. anterior-upper view of Sa element,
 982 GSC138376; 19. anterior view of Pa element, GSC138377; 20. inner lateral view of Sb element,
 983 GSC138378; 21. posterior view of M element, GSC138379. **22–28. *Aspelundia* cf. *As.***
 984 ***borenorensis* (Bischoff)** ($\times 60$); from 469 (except 25 from 478), section 12; 22. anterior view of
 985 Pa element, GSC138380; 23. posterior view of Sb element, GSC138381; 24. inner lateral view of
 986 Sc element, GSC138382; 25. inner lateral view of Pb element, GSC138383; 26. posterior view of
 987 M_1 element, GSC138384; 27. upper-anterior view of Sa element, GSC138385; 28. posterior view
 988 of M_2 element, GSC138386. **29. *Ozarkodina confluens* (Branson and Mehl)** ($\times 60$); from 696,
 989 section B; lateral views of Pa element, GSC138387. **30, 32–34. *Ctenognathodus* sp.** ($\times 60$); from
 990 696 (except 32 from 671), section B; 30. Lateral view of Pa element, GSC 138388; 32. posterior
 991 view of Sb element, GSC138390; 33. posterior view of Sa element, GSC138391; 34. inner lateral
 992 view of Sc element, GSC138392. **31. *Ozarkodina* sp.** ($\times 60$); from 696, section B; lateral views
 993 of Pa element, GSC138389. **35–36. *Ozarkodina parahassi* (Zhou, Zhai and Xian)** ($\times 70$); from
 994 525, section B; 35. lateral view of Pa element, GSC138393; 36. lateral view of M element,
 995 GSC138394. **37–40. *Wurmiella e. excavata* (Branson and Mehl)** ($\times 55$); from 493, section 12;
 996 37. inner lateral view of Sc element, GSC138395; 38. posterior view of Sb element, GSC138396;
 997 39. outer lateral view of Pb element, GSC138397; 40. outer lateral view of Pa element 138398,
 998 GSC; **41. *Kockelella?* sp.** ($\times 55$); from 41 from 601, section B; posterior view of M element,
 999 GSC138399. **42. *Ozarkodina* cf. *O. crista* (Walliser)** ($\times 100$); from 130, section 3; upper view of
 1000 Pa element, GSC138400. **43. *Ozarkodina polinclinata* (Nicoll and Rexroad)** ($\times 60$); from 413,
 1001 section B; lateral view of Pa element, GSC138401.

1002 **Fig. 7.** Silurian conodonts. **1–7. *Astropentagnathus irregularis* Mostler** ($\times 50$); 1, 3 and 7 from
 1003 440, section B; 2, 4, 5 and 6 from 129, section 3; 1. outer lateral view of Sc element, GSC138402;

1004 2. outer lateral view of Pb element, GSC138403; 3. posterior view of Sb element, GSC138404; 4.
 1005 posterior view of Sa element, GSC138405; 5. lateral view of M element, GSC138406; 6. upper
 1006 view of Pa₁ element, GSC138407; 7. upper view of Pa₂ element, GSC138408. **8. *Kockelella v.***
 1007 ***variabilis* Walliser** (×25); from 775, section B; upper view of Pa element, GSC138409. **9–10.**
 1008 ***Apsidognathus tuberculatus lobatus* Bischoff** (9 ×50; 10 ×40); 9 from 129, section 3; 10 from
 1009 471, section B; 9. upper view of arched stelliscaphate element, GSC138410; 10. upper view of
 1010 Pa element, GSC138411. **11. *Astropentagnathus* sp.** (×45); from 143, section 2; upper view of
 1011 Pa element, GSC138412. **12. *Aulacognathus?* sp.** (×25); from 477, section 12; upper view of Pa
 1012 element, GSC138413. **13. *Apsidognathus t. tuberculatus* Walliser** (×55); from 456, section B;
 1013 upper view of Pa element, GSC138414. **14–15. *Kockelella? trifurcata* Zhang and Barnes** (×70);
 1014 from 493, section 12; outer lateral and upper view of Pa element, GSC138415. **16.**
 1015 ***Aulacognathus angulatus* Bischoff** (×50); from 143, section 2; upper view of Pa element,
 1016 GSC138416; **17–18. *Aulacognathus bullatus* (Nicoll and Rexroad)** (×50); 17 from 413, section
 1017 B and 18 from 144, section 2; upper views of Pa element, GSC138417; 138418. **19–21.**
 1018 ***Kockelella patula* Walliser** (×25); 19 from 497, 20 from 489 and 21 from 493, section 12; 19.
 1019 inner lateral view of Sc element, GSC138419; 20. posterior view of Sa element, GSC138420; 21.
 1020 upper view of Pa element, GSC138421. **22–31. *Pterospathodus celloni* Walliser** (×60, except
 1021 27×50); 22–26 from 143, section 2; 27 from 130b, and 28 and 29 from 144, section 3; 30 and 31
 1022 from 440 and 456, section B; 22. outer lateral view of Sb element, GSC138422; 23. outer lateral
 1023 view of M element, GSC138423; 24. outer lateral view of Sc element, GSC138424; 25 and 29.
 1024 outer lateral view of Pb₁ element, GSC138425, 138429; 30. outer lateral view of carnuliform
 1025 element, GSC138430; 26, 27 and 28. lateral view of Pa element, GSC138426, 138427, 138428;
 1026 31. outer lateral view of Pb₂ element, GSC138431. **32–33. *Pterospathodus eopennatus* Männik**

1027 (×66); from 129, section 3; inner and upper views of Pa (Morph 3) element, GSC138432. **34–38.**
 1028 ***Pterospathodus pennatus procerus* (Walliser)** (×100 except 35 and 37×70); 35 from 469, 37
 1029 from 479, and 34, 36 and 38 from 480, section 12; 34. outer lateral view of Pb element,
 1030 GSC138485; 35. outer lateral view of S (?) element, GSC138486; 36–38. upper views of Pa
 1031 element, GSC138487, 138488, 138489. **39–40. *Rhipidognathus?* sp.** (×60); from 226, section 5;
 1032 32. posterior view of Sa element, GSC138433; 33. lateral view of S element, GSC138434. **41–42.**
 1033 ***Kockelella? manitoulinensis* (Pollock, Rexroad and Nicoll)** (×55); from 130b, section 3; inner
 1034 lateral and upper views of Pa element, GSC138435.
 1035 **Fig. 8.** Silurian conodonts (1–16) and conodonts present in both Ordovician and Silurian strata
 1036 (17–32). **1–3. *Dapsilodus* sp.** (×55); 1 from 413, 2 from 367, and 3 from 671, section B; 1.
 1037 lateral view of M element, GSC138437; 2. lateral view of Sa element, GSC138438; 3. lateral
 1038 view of Sb-c element, GSC138439. **4–6. *Pseudobelodella spatha* (Zhou, Zhai and Xian)**
 1039 (×100); from 130a, section 3; 4. lateral view of acostiform element, GSC138440; 5. lateral view
 1040 of bicostiform element, GSC138441; 6. lateral view of unicostiform element, GSC138442. **7.**
 1041 ***Pseudooneotodus bicornis* Drygant** (×90); from 601, section B; upper view, GSC138443. **8–12.**
 1042 ***Walliserodus* cf. *W. sancticlairei* Cooper** (×75); 8 and 9 from 130a, section 3 and 10–12 from
 1043 145, section 2; 8. outer lateral view of unicostatiform element, GSC138444; 9. inner lateral view
 1044 of curvatiform element, GSC138445; 10. outer lateral view of debolotiform element,
 1045 GSC138446; 11. lateral view of dyscritiform element, GSC138447; 12. inner lateral view of
 1046 debolotiform element, GSC138448. **13–15. *Decoriconus fragilis* (Branson and Mehl)** (×90);
 1047 from 146, section 2; 13. inner lateral view of acontiodontiform element, GSC138449; 14, inner
 1048 lateral view of drepanodontiform element, GSC138450; 15. inner lateral view of paltodontiform
 1049 element, GSC138451. **16. ?*Dentacodina dubia* (Rhodes)** (×60); from 130a, section 3; lateral

view of denticulate element, GSC138452. **17–19. *Walliserodus curvatus* (Branson and Branson)** ($\times 65$ except 17×50); from 145, section 2; 17. inner lateral view of deboltiform element, GSC138453; 18. lateral view of dyscritiform element, GSC138454; 19. outer lateral view of unicostatiform element, GSC138455. **20–24. *Panderodus recurvatus* (Rhodes)** ($\times 65$); from 451, section 13; 20. inner lateral view of arcuatiform element, GSC138456; 21. lateral view of aequaliform element, GSC138457; 22. inner lateral view of compressiform element, GSC138458; 23. inner lateral view of tortiform element, GSC138459; 24. inner lateral view of asymmetrical graciliform element, GSC138460. **25–31. *Panderodus unicostatus* (Branson and Mehl)** ($\times 55$); from 130, section B; 25. subsymmetrical graciliform element, GSC138461; 26. inner lateral view of arcuatiform element, GSC138462; 27. lateral view of aequaliform element, GSC138463; 28. inner lateral view of truncatiform element, GSC138464; 29. inner lateral view of tortiform element, GSC138465; 30. outer lateral view of asymmetrical graciliform element, GSC138466; 31. inner lateral view of compressiform element, GSC138467. **32. *Pseudooneotodus beckmanni* (Bischoff and Sannemann)** ($\times 90$); from 451, section 13; upper view, GSC138468.

Fig. 9. Upper Ordovician and Silurian stratigraphy on Grinnell Peninsula, Devon Island and southern Ellesmere Island, and its correlation with the Geological Time Scale (GTA) 2012. The Upper Ordovician GTS is from Cooper and Sadler (2012) and Silurian GTS is from Melchin et al. (2012). The dashed lines in the Conodont Zonation (GST 2012) denote uncertainty in the placement of that boundary with respect to the Stage slice. The dashed lines in the Conodont Zones (this study) denote uncertainty in the placement of that boundary with respect to both Stage slice and studied sections.

1072 **Fig. 10.** Conodonts with different CAI values. 1. Sb-c element *Dapsilodus* sp. (CAI=1), from
 1073 497, section 12, GSC138469; 2. compressiform element of *Panderodus unicostatus* (CAI=1),
 1074 from 499, section 12, GSC138470; 3. unicostatiform element of *Walliserodus curvatus* (CAI=1),
 1075 from 468, section 14, GSC138471; 4 and 5. oistodiform element of *Drepanoistodus suberectus*
 1076 (CAI=1.5–2), 4 from 451, section 13 and 5 from 213, section 5, GSC138472, GSC138473; 6.
 1077 compressiform element *P. unicostatus* (CAI=3), from 214, section 5, GSC138474; 7 and 8.
 1078 compressiform element of *P. unicostatus* (CAI=4), from 99 and 101, section 1, respectively,
 1079 GSC138475, GSC138476; 9. curvatiform element of *W. curvatus* (CAI=4), from 143, section 2,
 1080 GSC138477; 10. Pa element of *Astropentagnathus irregularis* (CAI=5), from 129 section 3,
 1081 GSC138478; 11. compressiform element of *P. recurvatus* (CAI=4), from 130b, section 3,
 1082 GSC138479; 12. arcuatiform element of *P. unicostatus* (CAI=5), from 577, section B,
 1083 GSC138480; 13. compressiform element of *P. recurvatus* (CAI=4), from 0, section B,
 1084 GSC138481; 14 and 15. arcuatiform element of *P. recurvatus* (15, bottom view showing basal
 1085 filling being replaced by bitumen) (CAI=6.5), from 374, section B, GSC138482; 16 and 17.
 1086 compressiform element *P. unicostatus* (16, inner view of 17) (CAI=6.5), from 577, section B,
 1087 GSC138483; 18. dyscritiform element of *W. cf. W. sancticlairei* (CAI=6.5), from 374, section B,
 1088 GSC138484. White scale bar at bottom right is for all images except for 10.

Figure 1
 173x182mm (300 x 300 DPI)

Figure 3
156x212mm (300 x 300 DPI)

Figure 4
225x173mm (300 x 300 DPI)

Figure 5
175x204mm (300 x 300 DPI)

Figure 6
177x230mm (300 x 300 DPI)

Figure 7
175x237mm (300 x 300 DPI)

Figure 8
175x144mm (300 x 300 DPI)

Years (Ma)	Period	Epoch/Age/Stage/Stage slice		Conodont Zonation (GTS 2012)		Conodont Zones (this study)	Lithostratigraphy	
							shelf facies	basin facies
425 <								

Figure 9
164x202mm (300 x 300 DPI)

Figure 10
165x135mm (300 x 300 DPI)