

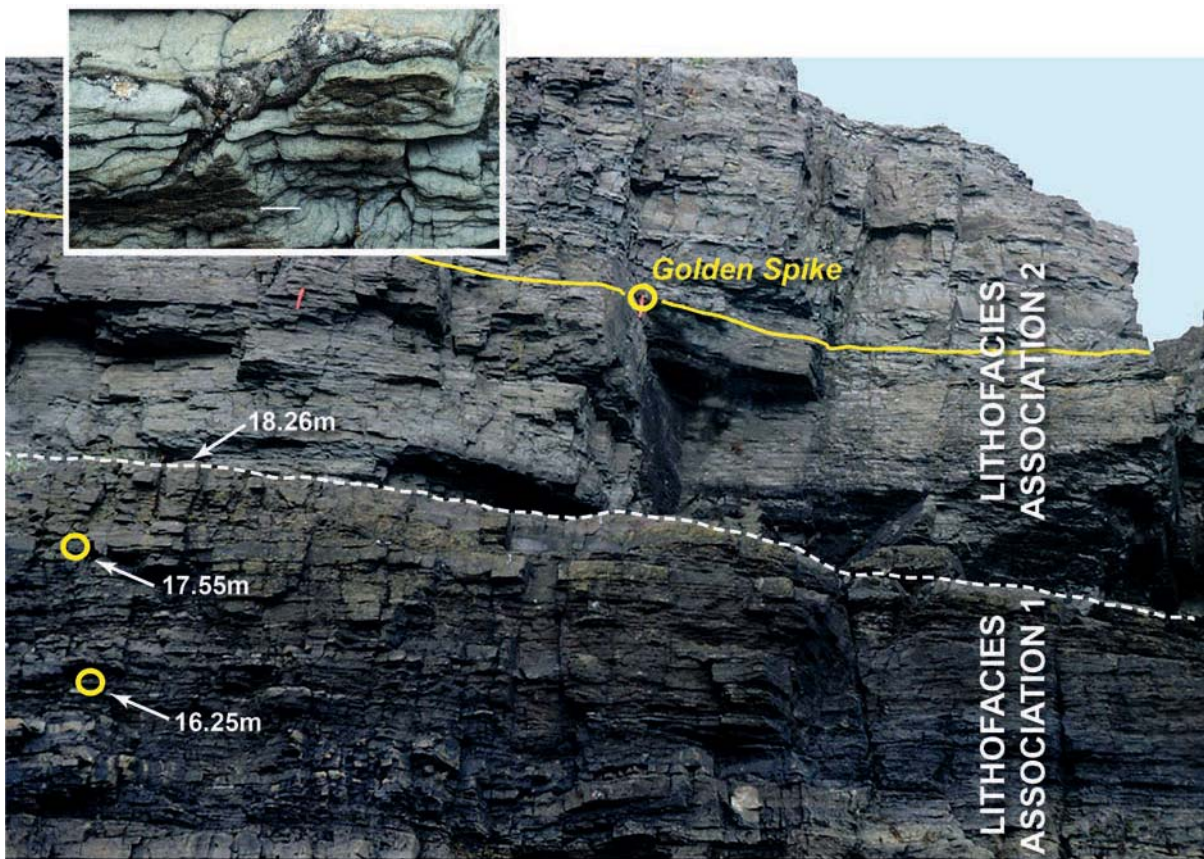


Ediacaran–Cambrian of Avalonian Eastern Newfoundland

(Avalon, Burin, and Bonavista Peninsulas):

International Symposium on the Ediacaran–Cambrian Transition,

FIELD TRIP 4



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LEADERS: Ed Landing, Paul M. Myrow, Guy M. Narbonne, Gerd Geyer, Luis A. Buatois, M. Gabriela Mángano, Alan Kauffman, Stephen R. Westrop, Björn Kröger, Brittany Laing, and Romain Gougeon

COVER

Ediacaran–Cambrian boundary global standard section and point (GSSP) at Fortune Head, Burin Peninsula, southeastern Newfoundland marked by designation “Golden Spike” 2.4 m above the base of lithofacies association 2/traditional member 2 of the lower Chapel Island Formation (compare local stratigraphic terminology at Stops 4–6 by Landing, 1996; Landing and Westrop, 1998a, b; Geyer and Landing, 2017 with Narbonne et al., 1987; Myrow et al., 2012). The GSSP lies in the regionally extensive (SE New Brunswick–Burin Peninsula, SE Newfoundland) lower lithostratigraphic member (Quaco Road Member) of the Chapel Island Formation (Landing, 1996; Landing and Westrop, 1998a, b; Geyer and Landing 2017). The GSSP horizon is best defined at the base of the Treptichnus pedum Assemblage Zone within the lower range of T. pedum and immediately above the highest occurrence of the Ediacaran taxa Harlaniella and Palaeopascichnus (see Narbonne et al., 1987; Landing et al., 2013b). Inset shows T. pedum specimen with burrow probes at the GSSP horizon (at 21 m in the measured Fortune Head section); horizontal white line is 1.0 cm. Pictures by G. Geyer in 1987.

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St. John's, Newfoundland, Canada

2017

sponsored by

The International Subcommittee on Ediacaran Stratigraphy (ICES)

The International Subcommittee on Cambrian Stratigraphy (ISCS)

June 15–29, 2017

St. John's, Newfoundland, Canada



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Recommended citation

Landing, E., Myrow, P.M., Narbonne, G.M., Geyer, G., Buatois, L.A., Mángano, M.G., Kaufman, A.J., Westrop, S.R., Kröger, B., Laing, B. and Gougeon, R., 2017, Ediacaran–Cambrian of Avalonian Eastern Newfoundland (Avalon, Burin, and Bonavista Peninsulas), Landing, E., Myrow, P.M., Geyer, G., McIlroy, D., eds., International Symposium on the Ediacaran–Cambrian Transition: Field Trip 4.

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INTRODUCTION

Geological Significance of Eastern Newfoundland (EL)

Avalonia and History of Geological Research

Pre-plate tectonic geological syntheses led to an early appreciation of the lithologic and biotic similarities of a belt of rocks in the northeast Appalachians and Britain. Walcott (1980, p. 567, 568) noted that the closest faunal and lithologic similarities of the lower Cambrian (his “*Olenellus Zone*”) of the “Atlantic Coast Province” was between SE Newfoundland, Wales, and NW England. These regions were later crucial to the application of early plate tectonic theory with recognition that inliers in the northeast Appalachians and southern Britain showed regional similarities in a late Precambrian, volcanic-rich basement and a lower Paleozoic siliciclastic-dominated cover succession with a distinctive biotic sequence. This “Avalon Platform,” named for its characteristic development in the Avalon Peninsula, SE Newfoundland (Williams, 1964; see also Rast *et al.*, 1976), is now regarded as an exotic terrane in the Acadian Mountains within the central Appalachians that extends from Rhode Island and eastern Massachusetts through coastal Maine and the central Maritime Provinces to SE Newfoundland (Figure 1). This terrane and the microcontinent it records (discussed below) have been termed “Avalonia” (Pickering *et al.*, 1988).

The eastern continuation of Avalonia was broken away with the Mesozoic opening of the Atlantic Ocean. This eastern region comprises much of Caledonian Wales and central England and continues through Belgium into western Germany (e.g., Landing and Westrop, 1998a; Linnemann *et al.*, 2008).

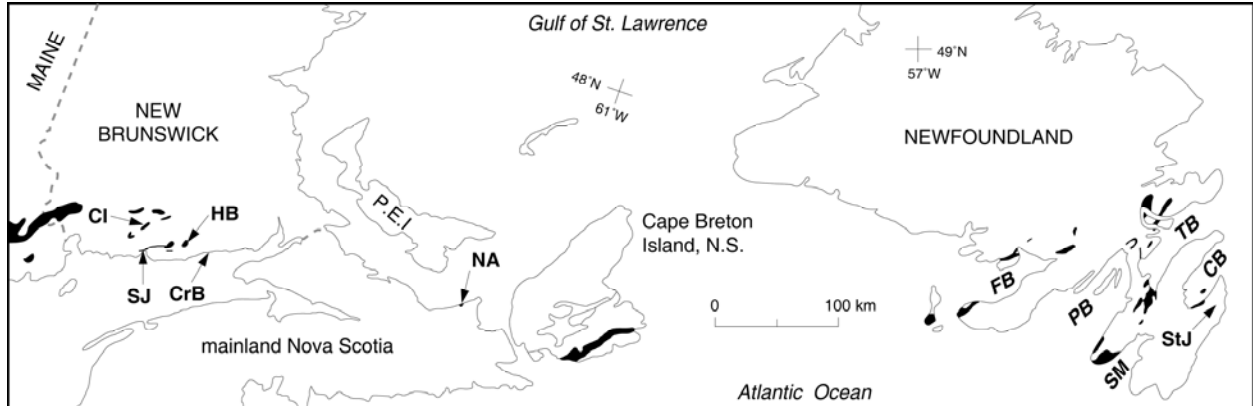


Figure 1. Inliers of terminal Ediacaran–Lower Paleozoic cover sequence (in black) in Avalonian Maine, Maritime Canada, and southeast Newfoundland. Abbreviations: CB, Conception Bay; CI, Caton’s Island; CrB, Cradle Brook; FB, Fortune Bay; HB, Hanford Brook; NA, Northern Antigonish Highlands; PB, Placentia Bay; P.E.I., Prince Edward Island; SJ, Saint John; StJ, St. John’s; TB, Trinity Bay. Modified from Landing (1996a, fig. 1).

Early work on Cambrian faunas in SE Newfoundland (e.g., Billings, 1865, 1872) included what proved to be the first description of a Precambrian animal fossil, *Aspidella terranova* Billings, 1872). Continuing research has meant that this easily accessible, well exposed area (i.e., shoreline and sea cliffs) has become an increasingly important succession for understanding Earth and biotic history. The uppermost Precambrian (Ediacaran System) of SE Newfoundland preserves a remarkable succession of Earth’s oldest large metazoans (Field trips 1, 3, 5 of this conference). Associated volcanic ashes allow a precise U-Pb zircon geochronology of the oldest known Ediacaran organisms and the preceding Gaskiers glaciation (Pu *et al.*, 2016, and references therein). Stops 2 and 3 (this Field Trip; Figure 2, localities SpB and UIC) allow examination of this older middle Ediacaran stratigraphy and biota.

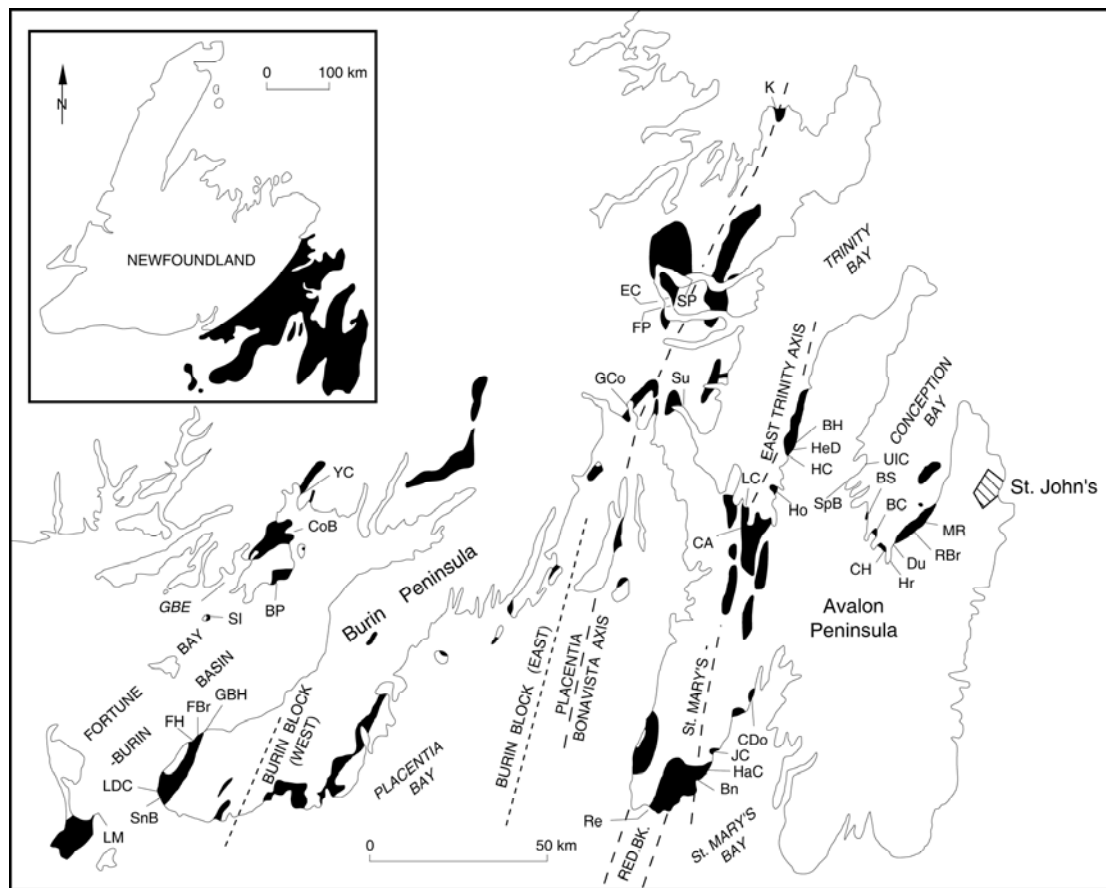


Figure 2. Terminal Ediacaran–Lower Ordovician inliers (black) in Avalonian southeastern Newfoundland, and localities for Field Trip 4 stops (see text, “Locality”). Abbreviations: BC, Bacon Cove; BH, Boar’s Head; Bn, Branch; BP, Blue Pinon Cove; BS, Brigus Point South; CA, Chapel’s Arm; CDo, Cape Dog; CH, Chapel’s Head; CoB, Corbin Bay; Du, Duffs; EC, Elliot Cove; FBr, Fortune Brook; FH, Fortune Head; FP, Fosters Point; GBE, Grand Bank de l’Eau; GCo, Goose Cove; HaC, Hay Cove; HC, Highland Cove; HeD, Heart’s Delight; Ho, Hopeall Head; Hr, Holyrood; JC, Jigging Cove; K, Keels; LC, Long Cove; LDC, Little Dantzic Cove; and Little Dantzic Cove Brook; LM, Langlade; MPE, McLeod Point; MR, Manuels River; RBr, Red Brook Road; Re, Redlands Cove; RED.BK., Redlands Block; SI, Sagona Island; SnB, Snook’s Brook; SP, Smith Point; SpB, Spaniard’s Bay; Su, Sunnyside; UIC, Upper Island Cove; YC, Young’s Cove. Figure modified from Landing and Westrop (1998, fig. 1).

Purposes of Field Trip

More recent work on the terminal Ediacaran–early Cambrian faunas of SE Newfoundland and the rest of Avalonia has led to a “fleshing out” of the Cambrian Evolutionary Radiation (CER) by documenting the driving forces that led to the origin, habitats of diversification, ecologic roles, and sclerotome construction of the oldest multicellular burrowing and biomineralized metazoans. Thus, a number of existing syntheses suggested by Avalonian work will be noted in the course of Field Trip 4:

1) that the shallow-water origin of deep burrowing and skeletalized taxa suggested in Avalonian sections may reflect metazoan “discovery” of largely uninhabited peri/intertidal environments subject to high UV-B and daily desiccation (Landing and Westrop, 2004; Meert *et al.*, 2016);

2) that the absence of early Cambrian tropical platforms, evaporites, and thrombolites; any archaeocyaths; and a very late appearance of trilobites (without fallotaspids) comport with Avalonia being a middle latitude continent distant from equatorial Gondwana by the terminal Ediacaran (Landing, 1992a, 1996a; 2005; Landing *et al.*, 2013b);

3) that, as in modern oceans, the dominate biomineral was aragonite, not phosphatic minerals, as first noted in early Cambrian fossil assemblages of Avalonia (Landing *et al.*, 1989; Landing, 1992a);

4) that Avalonian fossil assemblages supplied key information on ecologic roles and skeletal construction CER taxa with (a) hyoliths not limited to “shell draggers” but included forms that lived like scaphopods (Landing and Kröger, 2012), that (b) the earliest mid-water macropredators were pseudoconodonts (e.g., Landing, 1991), and (c) all tommotiid groups were

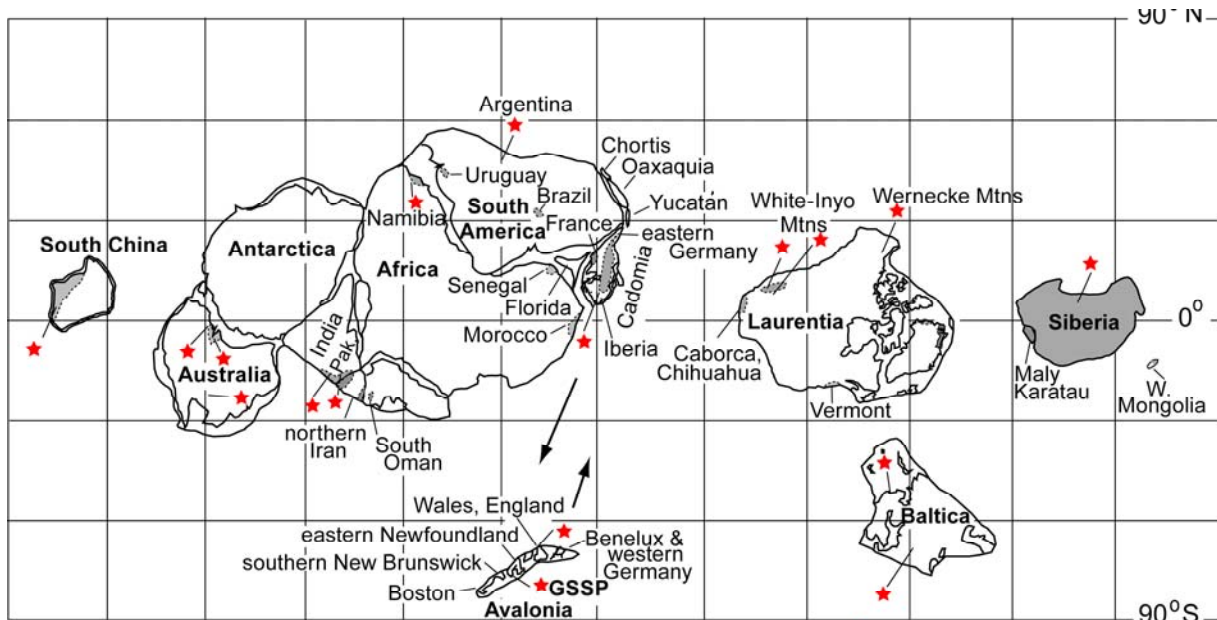
early shown to have closely-spaced, often ontogenetically fused phosphatic sclerites in Avalonian collections (Landing, 1988, 1995);

5) that there was a ca. 10 m.y. “stasis” of early, mollusk-rich, Stage 2/Laolinian Stage microfaunas, which were only later accompanied by the late origin of archaeocyaths (e.g., Landing and Kouchinsky, 2016); and

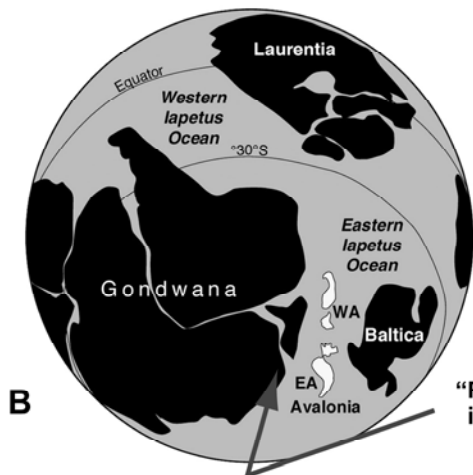
6) that volcanism associated with transcurrent, not rift, faulting along the terrane has allowed for numerous, precise, early Cambrian–Ordovician U-Pb zircon dates in North American and British Avalon. These U-Pb dates allow determination of rates of faunal replacements, accumulation rates of compacted rock, and improved interregional correlations (e.g., Davidek *et al.*, 1998; Landing *et al.*, 1998, 2000, 2013a; Harvey *et al.*, 2011; Williams *et al.*, 2013).

Generalized Terminal Ediacaran–Lower Paleozoic Deposition, Stratigraphy, and Biotas

A continuous succession does not exist through much of the Ediacaran into the terminal Ediacaran–Cambrian anywhere in Avalonia. Indeed, the Avalonian terminal Ediacaran–Cambrian overlies older Ediacaran strata as seen at Stops 2 and 3 (this Field Trip) with angular unconformity or nonconformably contacts on early Ediacaran intrusive rocks in SE Newfoundland (Stops 1, 8). This overlap succession in SE Newfoundland (Figure 3) is particularly important for terminal Ediacaran–Cambrian correlations and chronostratigraphy as it provides the global stratotype section and point (GSSP) and comparative sections for the Ediacaran–Cambrian boundary (Stops 4, 5; Narbonne *et al.*, 1987; Landing, 1992, 1994; Brasier *et al.*, 1994a; Landing *et al.*, 2013b; Geyer and Landing, 2016). The long-term utility of the GSSP for the Ediacaran–Cambrian boundary for global correlation (e.g., Landing *et al.*, 2013b,



A Basal Cambrian *Trichophycus pedum* Assemblage Zone assemblages (★) on revised terminal Ediacaran–Terreneuvian palaeogeographic map; = tropical carbonate platforms (+/- evaporites) & evaporitic successions (Hormuz Basin, South Oman, Salt Range, Siberian Olekma facies salt), = movement of Avalonia & West Gondwana through Early Cambrian



Traditional "Antarctocentric" palaeogeographic map for terminal Ediacaran through Terreneuvian; West Gondwana placed at or near South Pole and Avalonia not a unified palaeocontinent, from Dalziel (1997).

"Polar" evaporites, archaeocythans, and oosparites in southern Morocco

Figure 3. Terminal Ediacaran–lowest Cambrian paleogeography. (A) Revised paleogeographic map shows dominantly tropical distribution of major continents, preserved extent of tropical carbonate platform and evaporite-rich successions (light grey areas), and location of basal Cambrian *Treptichnus/Trichophycus pedum* Assemblage Zone ichnofossil assemblages. Figure 4A and 4B modified from Landing et al. (2013b, fig. 2). (B) Traditional early Cambrian paleogeographic map with major continents at high south latitudes emphasises mismatches between interpreted paleolatitudes and lithofacies and biotas of major continents. Abbreviations: EA and WA, east and west Avalonia, respectively. Selected sources for: *Trichophycus pedum* Assemblage Zone localities (Narbonne et al., 1987; Geyer and Uchman, 1995; Gamez Vintaned,

1996; Li *et al.*, 1996; Narbonne and MacNaughton, 1999; Gehling *et al.*, 2001; Geyer, 2005; Rogov *et al.*, 2016); numerous localities in South China (Weber *et al.*, 2007) shown by two arrows. Selected geological summaries: Avalonia (Landing, 1996a, 2004, 2005; Landing and Westrop, 1998a); Baltica (Nielsen and Schovsbo, 2011); Caborca and Chihuahua regions, northern Mexico (Sánchez-Zavala *et al.*, 1999, and sources therein); France (Doré, 1969; Pillola, 1993); Iran (Hamdi, 1995); Germany (Geyer *et al.*, 2008); Greenland (Cowie, 1971); Iberia (Liñán *et al.*, 1993, 2006); Maly Karatau, Kazakhstan (Missarzhevsky and Mambetov, 1981); Mongolia (Brasier *et al.*, 1996; Smith *et al.*, 2016; Landing and Kruse, 2017); Morocco (Geyer and Landing, 1995, 2006); Oman (Schröder *et al.*, 2007); Siberia (Brasier *et al.*, 1994b; Rozanov and Zhuravlev, 1996; Rogov *et al.*, 2016); South China (Steiner *et al.*, 2004, fig. 13); Uruguay and Brazil (Gaucher *et al.*, 2003, 2007); Vermont (Landing *et al.*, 2007; Landing, 2012); White-Inyo Mountains, California–Nevada (Mount and Signor, 1992; NW Canada (MacNaughton and Narbonne, 1999).

2015c) is enhanced by new studies of organic carbon stratigraphy and trace fossil biostratigraphy.

Important faunas and chemostratigraphic data through the lower part of the upper Terreneuvian Series (the proposed Laolinian Stage of Landing *et al.*, 2013b, or informal Stage 2, of the lowest Cambrian) are documented for the nearby Little Dantzic Cove section (Stop 6A). It is notable that in higher rocks, body fossils of Earth's oldest trilobites are absent in Avalonia. This means that intervals coeval with the oldest known trilobite-bearing strata of Siberia (Cambrian informal Series 2, Stage 3, or proposed Lenaldanian Series and Zhurinskyan Stage of Landing *et al.*, 2013b) occur in SE Newfoundland but lack trilobite fossils. Indeed, this interval comprises the sub-trilobitic, upper *Camenella baltica* Zone (Stops 10, 11B).

Trilobite fossils appear only higher across North American and British Avalonia, where they appear relatively late. Avalonian trilobites have their lowest occurrence relatively late in

informal Series 2, Stage 4 (upper Lenaldanian Series) with marine onlap across a trans-Avalonian unconformity (Stops 6B, 8, 10). The late appearance of Avalonian trilobites first demonstrated that the lowest appearance of trilobite fossils was significantly diachronous globally (Landing *et al.*, 1989, 2013b). The base of the oldest trilobites, the middle lower Cambrian boundary interval (informal Cambrian Stage 3–4), and the lower–middle Cambrian boundary interval (informal Stage 4–5) all lie at sequence boundaries in SE Newfoundland (Stops 8, 10 and Stop 7, respectively), and suggest an external control (eustatic or epeirogenic?) at these faunal turnover horizons.

Avalonia is regarded as an isolated terminal Ediacaran–Early Paleozoic continent (discussed below; Figure 3). The absence of archaeocyaths, a characteristic group of sponges in the early Cambrian, and their replacement in similar shallow-water habitats by “worm”-calcareous mud mound buildups (Stops 10, 11A) is consistent with the interpretation of Avalonia as a cool-water continent latitudinally distant from such tropical regions as the West Gondwana continent.

Terminal Ediacaran–Early Paleozoic Paleogeography of Avalonia (EL, GG)

Contrasting Paleogeographic Syntheses

Global biostratigraphic correlations and development of a stable chronostratigraphy through the Cambrian Evolutionary Radiation interval (i.e., terminal Ediacaran–early Cambrian) have been controversial. Problems arise from faunal provincialism and onshore–offshore litho- and biofacies linkages that reflect the latitudinal and longitudinal distribution and separation of Cambrian paleocontinents. Many existing paleogeographic maps do not resolve the pattern of terminal Ediacaran–early Cambrian biotic and lithofacies distributions of two key regions. These

two regions are Avalonia, as represented by the rocks and faunas of this field trip, and the northwest African–northern South American margin of West Gondwana (Figure 3). In part, contradictions between different terminal Ediacaran–Cambrian paleogeographic maps, the distribution of climatically-controlled lithofacies, and location of biotic provinces have arisen with the often low quality of paleomagnetic data [95% confidence limits commonly >20° and occasionally >40° and “acceptable” age uncertainties of about 10% on individual poles (Smith, 2001)].

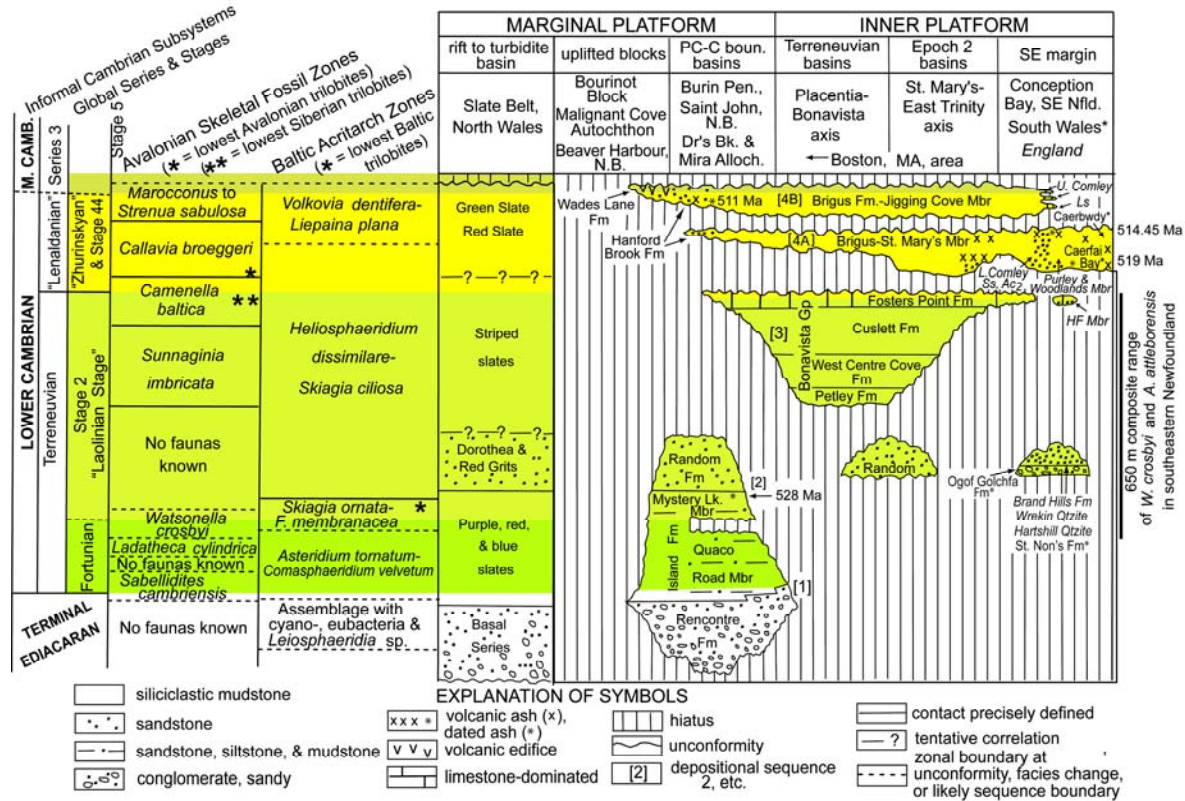


Figure 4. Terminal Ediacaran–lower middle Cambrian depositional sequence stratigraphy on marginal and inner platforms of the Avalon paleocontinent from England to eastern Massachusetts. Modified from Landing et al. (2013b, fig. 3). Fletcher’s (2003, 2006) alternative lithostratigraphic terms in SE Newfoundland are junior synonyms of Landing (1996a) and Landing and Westrop’s (1998b) units and frequently rely on non-lithologic and non-correlatable features (e.g., “separation planes”) in their definition (noted in Westrop and Landing, 2012).

Fletcher (2003, 2006) did not discuss or compare his new units with the earlier nomenclature, and proposed a number of lithostratigraphic names that were identical to the established nomenclature (e.g., his members A–D are Landing and Westrop’s (1998b) members 1–4 of the Manuels River Formation) or homonyms of the established nomenclature (e.g., his St. Mary’s Limestone, upper Chamberlain’s Brook Formation, and St. Mary’s Member, lower Brigus Formation, of Landing and Westrop, 1998b).

Paleogeographic reconstructions may also differ from each other as biogeographic changes or similarities between continents are not compared through geologic time in many reports. Thus, Fortey *et al.* (1996), Fortey and Cocks (2003), and Álvaro *et al.* (2013) all emphasized the similarity of lower–middle Cambrian boundary interval trilobite faunas as indicating Avalon–West Gondwana unity in the Cambrian–Early Ordovician. However, those reports did not discuss faunas and lithofacies from the first 20 m.y. of the Cambrian.

The earliest trilobite faunas from Avalonia and West Gondwana completely differ, and archaeocyaths are absent from Avalonia but form major buildups in supposedly adjacent West Gondwana (concluded to lie only “900 km” from SE Newfoundland; Fletcher, 2003). Furthermore, lower Cambrian shallow-water successions are siliciclastic-dominated in Avalonia (e.g., Myrow and Landing, 1992; Figure 4) and contrast with West Gondwana’s thick carbonate platforms (Geyer and Landing, 1995, 2006). These features indicate separation of Avalonia and West Gondwana by the terminal Ediacaran and not, as suggested (e.g., Fortey and Cocks, 1992, 2003), later in the Ordovician (Landing, 1996, 2005; Landing *et al.*, 2013a, b). Thus, the use of paleomagnetic data and an incomplete understanding of coeval lithostratigraphic successions has led to misfits between calculated paleolatitude, apparent polar wander paths, and the lithofacies and biotas of the two paleocontinents.

As detailed elsewhere (Landing, 1996; Landing and Westrop, 2004; Landing *et al.*, 2013a, b, 2015a), two contrasting paleogeographic models exist. These include “Antarctocentric” and “Fragmented Avalon” maps with the West Gondwanan margin at high south latitudes or at the South Pole and with Avalonia, often fragmented into tiny, isolated terranes, as part of, or close to, this margin (Figure 3B). Alternatively, the “Insular Avalon” model regards Avalonia as an elongate, higher latitude continent separate from tropical West Gondwana by the late Ediacaran. In this field trip guide, the latter paleogeographic reconstruction is utilized (Figure 3A) and does not regard Avalonia as close to or “peri-Gondwanan” by the terminal Ediacaran (*contra* Álvaro *et al.*, 2013, and references therein).

“Antarctocentric” and “Fragmented Avalon” Maps

Many terminal Ediacaran–early Cambrian maps are “Antarctocentric,” with the South American–West African–Cadoimian margin of West Gondwana at high latitudes or the South Pole. This interpretation of West Gondwana and with an Avalonian “appendage” began with Scotese *et al.* (1979) and Smith *et al.* (1981), who cited little data for their Cambrian–Early Ordovician maps. McKerrow *et al.*’s (1992) early Cambrian paleogeography is similar and based largely on biotic and interpreted paleoclimatic data and available paleomagnetic information. Repetition of these reconstructions with a seeming “southward march” of West Gondwana–Avalonia in successive publications, has led to an “Antarctocentric” paradigm for the West Gondwanan margin (e.g., Barr and Raeside, 1989; Fortey and Cocks, 1992, 2003; Keppie *et al.*, 1996; Torsvik *et al.*, 1996; Dalziel, 1997; Keppie and Ramos, 1999; Murphy *et al.*, 1999; Smith, 2001; Brasier and Lindsay, 2001; Fletcher, 2003, 2006; Álvaro *et al.*, 2003, 2013; Linnemann *et al.*, 2008; Parkhaev and Karlova, 2011; Zhuravlev *et al.*, 2012; Boucot *et al.*, 2013). One

exception is Piper (1987), who interpreted paleomagnetic data to reconstruct an Arctic Avalonia allied to Baltica.

“Antarctocentric” reconstructions (Figure 4A) place the locally evaporitic carbonate platforms of northwest West Gondwana (e.g., Morocco, Iberia, France, south-central Germany) and southwest Gondwana (Namibia, Uruguay, southern Brazil, Oman) at impossibly high and, thus, cool latitudes (Landing, 1996a, 2005, 2016; Figure 4B). This concern was discussed by Scott (2001), who noted that early Cambrian archaeocyaths and carbonate platforms were likely tropical and belied a 70° S–South Pole position of southern Morocco (*fide* Dalziel, 1997; Keppie and Ramos, 1999). Indeed, the locally 2.0 km-thick terminal Ediacaran–early Cambrian carbonate platform with local evaporites and overlying oolitic grainstones is wholly consistent with an equatorial–tropical location of southern Morocco and West Gondwana (Geyer and Landing, 1995, 2006; Landing, 1996a; Landing and Westrop, 2004, fig. 1). Álvaro *et al.* (2003) claimed to recognize carbonate platform development in the Comley area, NW England, to buttress an early Cambrian Avalonia–West Gondwana unity. However, the Comley limestone deposits (Figure 3) are only ca. 2.0 m thick (e.g., Cobbold, 1921, 1931; Rushton, 1974), and comprise a very thin, condensed succession of fossil tempestites that mark successive transgressive highstand deposits of a cool-water, siliciclastic-dominated, non-Gondwanan setting (Landing, 1996a; Harvey *et al.*, 2011).

A further caution for South America–NW Africa reconstructions at the South Pole or high southern latitudes (e.g., respectively Boucot *et al.*, 2013, and Fortey and Cocks, 2003; Álvaro *et al.*, 2013) is that low insolation of large polar continents led to “icehouse” conditions with continental ice sheets later in the Phanerozoic (Landing, 2016). However, the Cambrian had greenhouse or “hyperwarming” climates (Landing, 2012, and references therein). Similarly,

Dalziel's (1997; Figure 3B) map shows continents around an isolated south polar ocean, but this would have meant pack ice and, likely, adjacent continental ice sheets, as is the case with the modern Arctic Ocean. However, with exception of dropstone beds in Avalonian New Brunswick and southern Ireland, the terminal Ediacaran–Cambrian lacks evidence for significant glaciation (Landing and MacGabhann, 2010).

An “Antarctocentric” and “Fragmented Avalonia” interpretation with isolated Avalonian terranes located near West Gondwana has also been argued (Barr and Raeside, 1989; Dalziel, 1997; Figure 3B). Indeed, Steiner *et al.*'s (2007, fig. 13) map shows an “east” and “west” Avalonia as widely separated “bumps” on opposite (west South American and west African) margins of Gondwana.

Insular Avalon

As noted above, the Avalonian inliers in the Acadian–Caledonian orogen show remarkably similar terminal Ediacaran–Lower Paleozoic successions of siliciclastic-dominated lithologic units; faunal assemblages; and depositional sequence architecture (e.g., Landing, 1996a; Rees *et al.*, 2014; Figures 2, 3). Distinctive lithostratigraphic patterns (Figure 4), including (1) regional appearance of a thick feldspathic quartz arenite tidalite formation in the lower lower Cambrian or Terreneuvian Series (i.e., the Random Formation at Stops 6A, 6B, 8, 9; e.g., Landing *et al.*, 2013a, b), (2) a red limestone unit unconformably overlain by the lowest trilobite-bearing rocks that extends from Massachusetts to England (Fosters Point Formation, Stops 8, 10), and (3) a regional change to persistently dysoxic black mudstone deposition in the middle middle Cambrian or Drumian Stage (Landing, 1996a; Figures 3, 5, Manuels River

Formation), show that it is not credible to regard Avalonia as a collage of unrelated terranes in the early Paleozoic (Landing, 1996a; Landing *et al.*, 2013a, b; Figure 3).

Based on these data, Landing (1996a) proposed that Avalonia was a single early Paleozoic tectonostratigraphic unit (not Fortey and Cocks, 2003, *fide* Álvaro *et al.*, 2013), and that it had originated by the terminal Ediacaran (Figure 3A). Significantly, a uniform transtensional environment extended along the length of Avalonia in the terminal Ediacaran–Cambrian and had a role in defining the regionally similar stratigraphic succession along Avalonia (discussed below; Figure 4).

Generalized Terminal Ediacaran–Cambrian Stratigraphy (EL, SRW)

Epeirogenic Controls and Marginal–Inner platform

The terminal Ediacaran–lower Paleozoic of Avalonia unconformably overlies all horizons of a geologically complex, late Cryogenian–middle Ediacaran “basement” as a result of ca. 10 km of differential erosion (Landing and Benus, 1988b). This basement in North American Avalonia features a 200 m.y. record of extensional and arc-related tectonism. The upper part of this basement includes the fossiliferous middle Ediacaran (Field Trip Stops 2, 3 of this trip)—which is part of a volcanic-rich, upward-shoaling succession of continental slope to deltaic to subaerial facies (e.g., review by O’Brien *et al.*, 1996). This Avalonian basement has been considered to represent subduction–collisional or strike-slip regimes during a “Pan-African orogeny” (e.g., O’Brien *et al.*, 1983) or has been related to the Amazonian margin of a polar Gondwana (Keppie and Ramos, 1999, fig. 5).

Avalonia likely originated with rifting from West Gondwana that was completed by the late Ediacaran. The result of increasing latitudinal separation means the terminal Ediacaran–early

Cambrian, shallow-water, siliciclastic-dominated successions of Avalonia differ completely from the coeval West Gondwanan carbonate platforms. The similarity of lithofacies along the length of Avalonia arose in a strike-slip or transtensional tectonic environment, and not a rift (extensional) setting, because an eastward, step-like migration of depocenters and elevated blocks took place as the terminal Ediacaran–Cambrian progressed (Figures 2, 5). Lateral changes in thickness and different ages of unconformities across Avalonia with this eastward migration of depocenters do not indicate simple subsidence along a rift (Landing, 1996a; see Steel and Gloppen, 1980).

Syn depositional, transtensional faulting and ages of the oldest cover sequence units in eastern Newfoundland and other Avalonian regions (New Brunswick, Cape Breton Island, Wales) allow recognition of a “marginal platform” along the NW margin of Avalonia and an “inner platform” to the SE (Landing, 1994, 1996a, b, 2004; Landing *et al.*, 2013b). Both the marginal and inner platform are associated with early and late Cambrian and Early Ordovician acidic volcanism and middle Cambrian basic volcanism (Landing, 1996b; Landing *et al.*, 1997, 1998, 2000, 2008; Davidek *et al.*, 1998; Harvey *et al.*, 2001; Williams *et al.*, 2013).

Stratigraphic Nomenclature

The formation-level stratigraphy of Avalonian SE Newfoundland was established by studies that led to a regional nomenclature by the 1960s (e.g., Hutchinson, 1962). On the other hand, some publications proposed relatively distinct, formation-level nomenclatures specific to local areas of SE Newfoundland, such as the Burin Peninsula (e.g., Strong *et al.*, 1983; O’Brien and Taylor, 1983). Somewhat later work in the Avalon, Bonavista, and Burin peninsulas and along the south coast of Newfoundland emphasized the presence of regionally extensive,

formation-level lithologic units and predictable lateral thickness changes of formations as early suggested by Hutchinson (1962, figs. 2–5) who outlined long, linear, NNE-trending depocenters (e.g., Figure 2). Similarities in terminal Ediacaran–Lower Ordovician stratigraphy in SE Newfoundland and other Avalonian shallow-marine successions elsewhere in the NE Appalachians (Figure 1) and southern Britain reflect regionally extensive, syndepositional and likely fault-defined basins and uplifts. In addition, regionally extensive marker beds were interpreted as reflecting a unified sequence stratigraphic history (as transgressive dysoxic green mudstones or highstand carbonates in the siliciclastic-dominated terminal Ediacaran–Cambrian of SW Newfoundland (Landing and Benus, 1988a, b; Landing *et al.*, 1989; Myrow and Landing, 1992; Landing, 1996a, b).

Landing and Benus (1988b; also Landing, 1996a) modified the definition of the traditional Bonavista Formation to include much of the thin (to ca. 12 m) “Smith Point Limestone” and designated it the “Bonavista Group.” The uppermost, trilobite hash-rich and stromatolitic bed (or beds) of the “Smith Point” was recognized to unconformably overlie the microbial carbonate mound-dominated lower “Smith Point”—with the latter assigned to a new Fosters Point Formation that composed the top of the Avalonian sub-trilobitic lower Cambrian (Figure 5). This meant that the “Smith Point” was abandoned, and the traditional “top” layers of the “Smith Point” regarded as a tempestite limestone (or beds) and/or referred to the trilobite-bearing Brigus Formation (Landing and Benus, 1988b; Landing, 1996b; Stops 8, 10).

In addition, the Brigus Formation was recognized to consist of a lower St. Mary’s Member with a condensed limestone cap and an upper Jigging Cove Member (Landing and Westrop, 1998b). The Jigging Cove and the upper St. Mary’s members were preserved following

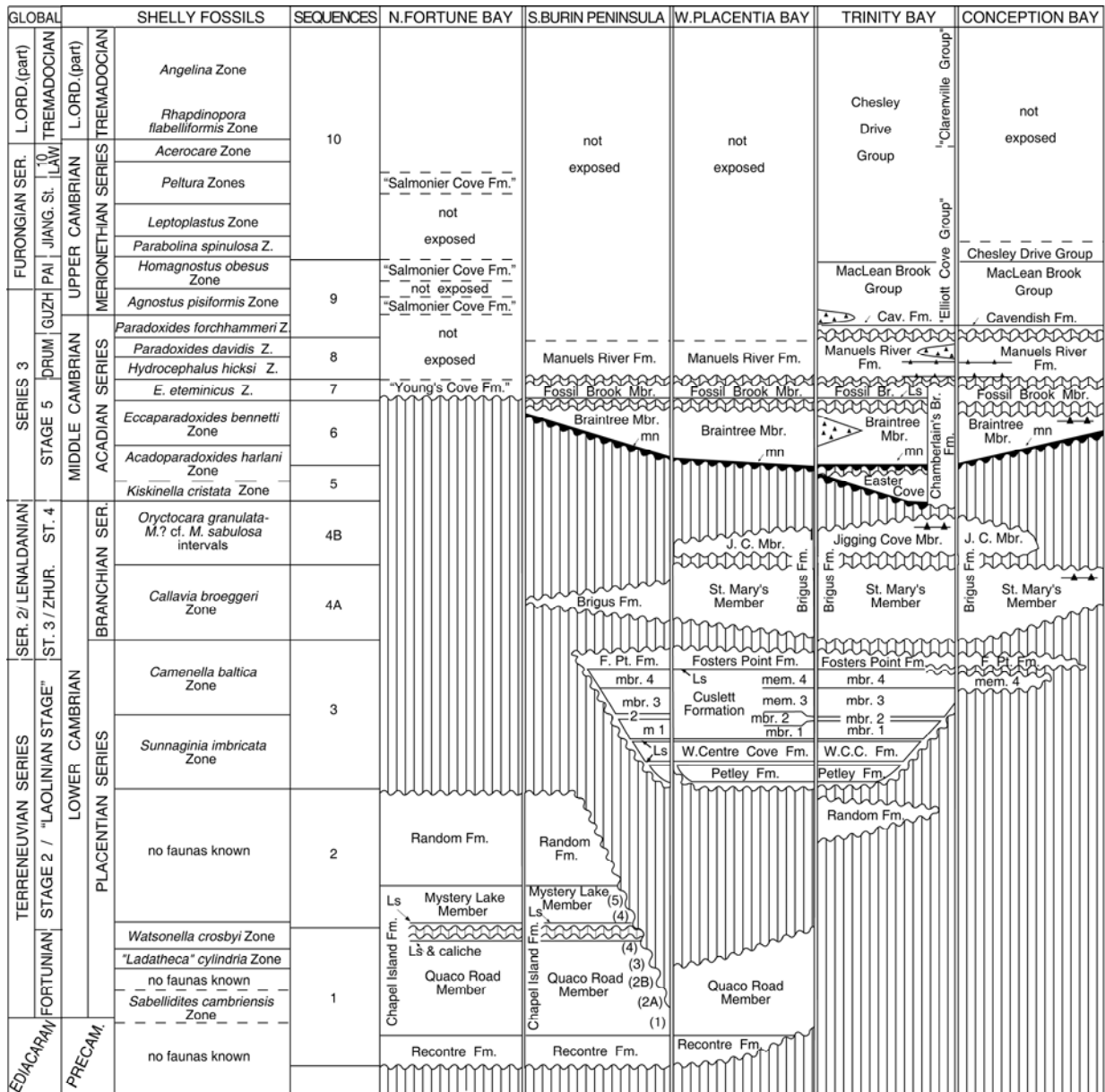


Figure 5. Epeirogenic depositional sequence stratigraphy in SE Newfoundland. Global chronostratigraphic divisions in two left columns; regional (Avalonian) Cambrian chronostratigraphic divisions, Placentian and Branchian stages, equivalent to Avalonian sub-trilobitic and trilobite-bearing lower Cambrian, respectively. Sub-trilobitic upper Camenella baltica Zone likely correlates with lower trilobite-bearing Series 2, Stage 3 (proposed Lenaldanian Series and Zhurinskyan Stage of Landing et al., 2013b) of Siberia. Base of Stage 2/proposed Laolinian Stage (Landing et al., 2013b) tentatively placed at peak of I' carbon isotope excursion (and equivalents) (see discussion in Landing and Kouchinsky, 2016;

Kouchinsky et al., 2017). Lawsonian Stage proposed by Landing et al. (2010; 2011). Avalonian Depositional Sequences 1–10 shown (see Landing, 1996a; Rees et al., 2014). Numbers in parentheses in Chapel Island Formation, e.g., “2B”, are earlier members of Chapel Island Formation, now designated as “lithofacies associations (Landing and Westrop, 1998a; Geyer and Landing, 2016). Symbols: vertical lines, hiatus; wavy line, unconformity; black triangles on lines, volcanic ashes; scattered black triangles, lenticular volcanic flows (=Cape Dog and Hopeall Head volcanics in upper Chamberlain’s Brook Fm., Hay Cove volcanics in Manuels River Fm., Chapel Arm Member volcanics in MacLean Brook Gp. Abbreviations: DRUM, Drummian Stage; F.Pt., Fosters Point; GUZH, Guzhangian Stage; J. C. M., Jigging Cove Member; Jiang, Jaingshanian Stage; LAW, Lawsonian Stage; Ls, Limestone; mn, manganese nodule bed; PAI, Paibaian Stage; SER., Series; ST., Stage; W. C. C., West Centre Cove; Z, Zone; 10, Stage 10. Modified from Landing and Westrop (1998a, figure 3).

pre-Chamberlain’s Brook Formation erosion only along the St. Mary’s-east Trinity axis (Landing and Westrop, 1998b; Figure 5; discussion of Stops 1, 7, 8).

The most reliable subdivision of the traditional lower middle Cambrian Chamberlain’s Brook Formation includes an originally unnamed lower, very manganese carbonate-rich member known only on the southern St. Mary’s-east Trinity axis (Landing, 1996b; Landing and Westrop, 1998b; Figures 2, 5). This lower member is best termed the Easter Cove Member (Fletcher, 2006; Stop 7). Higher Chamberlain’s Brook strata include a regionally extensive, much thicker Braintree Member and a thin, bedded limestone-rich, Fossil Brook Member cap in the St. Mary’s-east Trinity and eastern Placentia–Bonavista axes (Stops 1, 6B, 7).

Fletcher (2003) re-asserted the Bonavista Group and Smith Point Limestone without discussion or citation of the existing literature and named member-level subdivisions of the Brigus and Chamberlain’s Brook Formations. The 2003 report did not provide the required

criteria for proposing a lithostratigraphic unit (e.g., detailed lithology, measured sections, and type locality; see North American Stratigraphic Commission ([NAPC, below], 1983). Fletcher (2006) later provided these criteria, again without discussing the established lithostratigraphic divisions. Fletcher's (2006) procedure neither acknowledged that a stratigraphic nomenclature with priority over his existed (NAPC, 2005, Article 3c). Indeed, "stability of nomenclature is maintained by the rule of priority and by preservation of well-established names" and "priority of publication is to be respected" (NAPC, 2005, Article 3c). Fletcher (2003, fig. 3 caption) claimed that his lithostratigraphic units were "formally registered in the Canadian Lexicon of Stratigraphic Names ... in 1972," but this is puzzling as the 1985 Lexicon does not list any of his "named units" (Williams *et al.*, 1985). A fundamental criterion is that only formal publication in a peer-reviewed journal and not appearance in an unknown "formal registry" establishes a formal stratigraphic name (NAPC, 1983, 2005, Article 4). Fletcher (2010, unpublished communication to the International Subcommittee on Cambrian Stratigraphy) dismissed the Landing and Benus (1988b), Landing (1996), and Landing and Westrop (1998b) lithostratigraphic nomenclature as merely corresponding to depositional sequence stratigraphic units, while not noting that all of these earlier defined (1988b, 1996 and 1998b) units were defined lithostratigraphically and relied on bounding marker beds that happened to mark depositional sequence boundary intervals.

In any case, Westrop and Landing (2012) noted that NAPC instructions indicate that formations should correspond to genetic units and should not be split by major regional breaks, as Fletcher's (2006) "Smith Point Formation." It should be noted that among the marker beds not observed by Fletcher (2006, figs. 14, 15) is the thick limestone or laterally equivalent debris flow or glauconitic sandstone at the top of the St. Mary's Member (Landing and Westrop, 1998a, b; Stop 10). Fletcher (2003, p. 79) early regarded the Petley and West Centre Cove formations as

unmappable, presumably as he did not appreciate the regional extent of the bounding upper limestones of both formations. These latter limestones were later incorporated as the tops of the Petley and West Centre Cove “members” (Fletcher, 2006, fig. 11, modified without discussion from their proposed formation-level status and without acknowledgment of the nomenclature first proposed by Landing and Benus, 1988b, figs 35, 36). In this report, the lithostratigraphic terminology of Avalonian North America and the Avalonian depositional sequence stratigraphy defined by Landing and Benus (1988b), Landing (1996a), and Landing and Westrop (1998b) and recognized in South Wales (Rees *et al.*, 20014) are applied.

Marginal Platform

The marginal platform along the NW margin of Avalonia is defined by a thick, terminal Ediacaran–lowest Cambrian cover succession. This succession begins with a terminal Ediacaran conglomeratic unit (red beds and upper marginal marine facies) of the Rencontre Formation in Avalonian North America and comparable facies in southern Britain (Figures 3, 5). Myrow (1995) and Landing (1996a) provided sequence stratigraphic interpretation of the overlying Chapel Island Formation (Figures 5, 6). This includes a subsequent end-Ediacaran–early Cambrian Age 2/Laolinian Age deepening then shoaling of siliciclastic-dominated mudstone (lower–middle Chapel Island Formation) that is capped by a highstand carbonate of the middle Chapel Island in SE Newfoundland (Stops 4, 5) and a channel sandstone in southern New Brunswick (e.g., Landing and Westrop, 1998b; Landing and MacGabhann, 2010).

A Stage 2/Laolinian Stage depositional sequence boundary (Stop 6G) with an overlying, wave-dominated unit (upper Chapel Island Formation, member 5) followed by a tidalite sandstone (Random Formation) caps the terminal Ediacaran–lower Cambrian Stage 2 succession as Avalonian Depositional Sequence 2 (Figure 5). In SE Newfoundland, these strata extend

across the Fortune Bay Basin and the Burin Block, and onto the western Placentia–Bonavista Axis (Figures 3, 5; Landing, 1996a). Thin volcanic ashes exist in lower Stage 2 rocks in New Brunswick (e.g., Landing *et al.*, 1998).

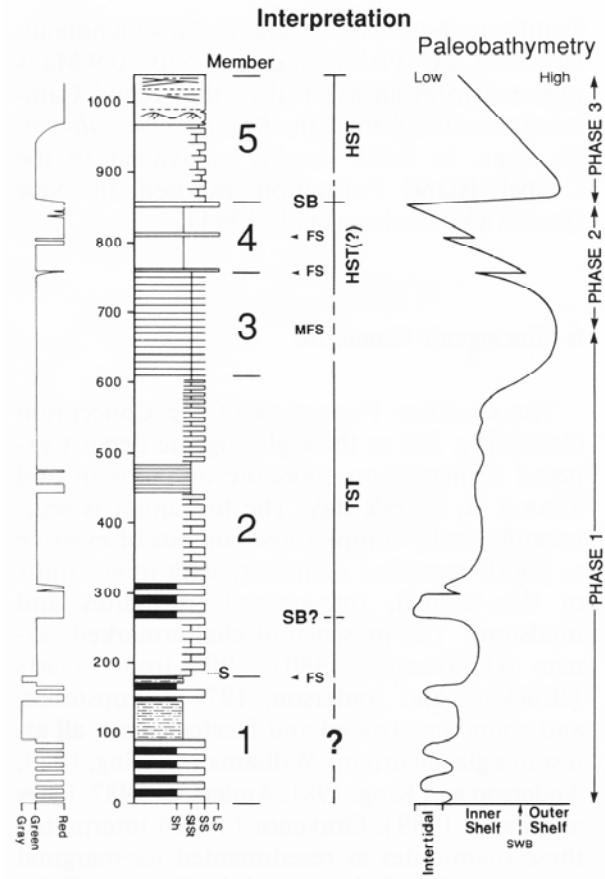


Figure 6. Generalized stratigraphic section for Chapel Island Formation showing inferred paleobathymetry, three-phase depositional history, and sequence-stratigraphic interpretation. Precambrian–Cambrian boundary stratotype shown by letter S. Informal members 1–5 of the Chapel Island Formation are equated to lithofacies associations 1–5 in this guide (also Geyer and Landing, 2016).

Chapel Island Formation Lithostratigraphic Nomenclature

As noted in the Figure 5 caption, this guide uses two alternative nomenclatural schemes for the lithostratigraphic subdivisions of the Chapel Island Formation. The traditional subdivision

of the Chapel Island Formation in the eastern Burin Peninsula has emphasized a vertical succession of five (1–5), informal, lithologically defined members of the Chapel Island Formation that comprise a generalized deepening–shoaling succession from the Rencontre Formation to the Random Formation (Bengtson and Fletcher, 1983; Myrow, 1987; Myrow *et al.*, 1988; Landing *et al.*, 1989; Myrow and Hiscott, 1993; Myrow *et al.*, 2012; Figures 6, 7).

Landing and colleagues (Landing, 1996a; Landing and Westrop, 1998a, b; Landing *et al.*, 2013a; Geyer and Landing, 2016; Stop 6A) recognize informal members 1–5. However, the latter reports have emphasized evidence for a subaerially eroded depositional sequence boundary within the uppermost limestone bed of member 4 (Stop 6A) to refer the lower Chapel Island to the Quaco Road Member and the very top of member 4 and all of member 5 to the Mystery Lake Member. (Both the Quaco Road and Mystery Lake members are regarded as regionally extensive on the marginal Avalonian platform from SE New Brunswick to Cape Breton Island and into the Burin Peninsula, and represent, respectively, the upper and lower parts of Avalonian Depositional Sequences 1 and 2 (Landing, 1996a; Rees *et al.*, 2014). Thus, the terms “members 1–5” and “lithofacies associations 1–5” are synonymous designations (e.g., Geyer and Landing, 2016). This guide attempts to always pair these designations (Figure 5).

Inner Platform

To the southeast, late onlap resulted in deposition of the Random Formation (tidalite sandstone) directly on Avalonian basement (Figure 5). This Random–Avalonian basement nonconformity or unconformity defines the Avalonian inner platform (Landing, 1996a, 2004). The inner platform extends from NW England and South Wales through SE Newfoundland (Avalon, Bonavista, St. Mary’s peninsulas) and to eastern Massachusetts and Rhode Island.

Epeirogenic uplift on the inner platform later led to partial or complete erosion of the tidalite sandstone (Figure 5; Stop 1), which was followed by subsistence of an elongate, fault-bounded block (Placentia–Bonavista Axis, “P–B A”; Figure 2). The P–B A received the greatest thicknesses of the sub-trilobitic Bonavista Group and its highstand carbonate cap (the Fosters Point Formation) (Figure 5). The green to red, siliciclastic mudstone-dominated Bonavista Group (upper Stage 2–Stage 3 or Laolinian–lower Zhurinskyan stages, see Figure 40) reaches 200 m in thickness in the P–B A (Stops 9–11B). The Bonavista Group thins by loss of its basal units with onlap both east and west of the P–B A (Landing and Benus, 1988a, b; Stop 8; Figure 5).

A subsequent eastward movement of Avalonian depocenters is recorded by subsidence along the St. Mary’s–Bonavista Axis (S–B A, Figure 2), with maximum accumulation of up to 300 m of the mudstone-dominated Brigus Formation (\approx Cambrian Stage 4) and lower Chamberlain’s Brook Formation (Stages 5–7; Stops 7, 10). These latter units onlap to the west and across the marginal platform, where they overlie Cambrian units as low as the Random Formation (Figure 5; Stop 6B).

On the inner platform, an offlap–onlap event separates lower Cambrian, upper Stage 4 units (St. Mary’s and Jigging Cove member of the Brigus Formation = Avalonian Depositional Sequences 4A–4B; Figures 3, 5; Stops 8, 10). Similarly, an offlap–onlap event and an unconformity in the lower–middle Cambrian boundary interval separates upper Cambrian Stage 4 from Stage 5 on the inner platform (i.e., upper Brigus and lower Chamberlain’s Brook Formation; Stop 7). Post-Stage 4 offlap may have involved epeirogenic uplift, as the Jigging Cove Member is absent east and west of the S–B A (Stops 1, 6; Figures 2, 5).

The fragmental limestone-rich Fossil Brook Member of the upper Chamberlain’s Brook Member (upper Stage 5) comprises the thin Avalonian Depositional Sequence 7 in Avalonian

New Brunswick and SE Newfoundland (e.g., Kim *et al.*, 2002; Figure 5). This sequence boundary is marked by reworked phosphatic grains and a burrowed firm-ground in SE Newfoundland (Stop 1), and by pre-fossilized and phosphatized skeletal debris eroded from the lower Cambrian Hanford Brook Formation in southern New Brunswick (EL, unpublished data).

A continuous, stratigraphically unbroken upper Chamberlain's Brook–lower Manuels River succession has been repeatedly asserted (Hutchinson, 1962, p. 23; Fletcher and Brückner, 1974; Fletcher, 2003, 2006). However, no evidence for stratigraphically unbroken sections through this formational contact exists anywhere in SE Newfoundland or southern New Brunswick (Landing, 1996a; Landing and Westrop, 1998a, b; Figure 5). Indeed, the differential loss of several meters of the upper Fossil Brook Member in the short distance between Manuels River and the Red Bridge Road areas (Landing and Benus, 1988a, b; Figure 2, localities MR and RB; Stop 1), suggests epeirogenic activity prior to onlap of the dysoxic mudstone deposits of the Manuels River Formation. A somewhat higher rate of subsidence along the St. Mary's-east Trinity axis is indicated by the doubling in thickness of the Manuels River Formation from Manuels River to the west side of St. Mary's Bay (e.g., Fletcher, 2006, fig. 23). The organic-rich Manuels River mudstones also onlap the marginal platform in the area of Stop 6 and indicate the complete submergence of SE Newfoundland, southern New Brunswick, and eastern Massachusetts–Rhode Island by dysoxic marine water (e.g., Skehan *et al.*, 1978; Landing, 1996a; Landing and Westrop, 1998a; Geyer and Landing, 2001).

Limited outcrop precludes a detailed sequence stratigraphic synthesis for strata above the Manuels River Formation. An unconformable upper contact of the Manuels River Formation indicates that the unit represents Avalonian depositional sequence 8 (Landing, 1996a; Landing and Westrop, 1998b; Figures 3, 5). Landing and Westrop (1998a, b) divided the thickest sections

of the Manuels River in eastern Trinity Bay and the SE St. Mary's Peninsula into successive informal members (1–4) of very dark, olive green-gray mudstone (members 1, 3) and black, pyritiferous mudstone (members 2, 4). Fletcher (2006) also recognized identical informal members (his A–D), without noting the earlier proposed members 1–4.

Regional marine offlap after Manuels River deposition is seen by local development of a thin stromatolitic limestone on the Manuels River (e.g., Landing and Westrop, 1998a, Highland Cove stop; Figure 2, locality HC). This offlap–onlap-associated unconformity (Avalonian depositional sequence 8–9 contact; Figures 3, 5) is marked by deposition of the southerly thickening, lower glauconitic sandstone of the Cavendish Formation (Landing and Westrop, 1998b). This offlap–onlap event marks a stage boundary with Drumian Stage trilobites (*Ptychagnostus punctuosus* Zone) in the upper Manuels River. Interestingly, we have now determined that *Paradoxides davidis* occurs in the feather-edge of the Cavendish Formation at Red Bridge Road (Figure 9) and this shows that the Avalonian Depositional Sequence 8–9 unconformity lies within the upper Drumian Stage—meaning that the duration of the offlap–onlap event was either brief or that the Drumian Age was unexpectedly long.

Guzhangian Stage (*Lejopyge laevigata* Zone) trilobites occur immediately above the thin Cavendish Formation at Manuels River (Landing and Westrop, 1998b, p. 28). The sandy strata of the Cavendish Formation range from a feather edge in the eastern St. Mary's-east Trinity axis (Stop 1) to ca. 45 m in western St. Mary's Bay (Landing and Westrop, 1998b; see Fletcher, 2003, 2006, his “Beckford Head Formation”).

FIELD WORKSHOP

Stop 1. Red Bridge Road (June 23; EL, SWR, GG)

Location

The section is SE of Saint John's and S of Kelligrews village (Figure 2, locality RBr). It is reached by turning west from Route 60 and traveling ca. 7 km south on Red Bridge Road. The gently east-dipping section (Figure 7) is almost completely exposed in a series of shallow road metal quarries that extend for about 300 m on the north side of Red Bridge Road and immediately north of a soccer field.

Generalized Stratigraphy and Significance

This inner platform succession on the far eastern limb of the St. Mary's–east Trinity axis (depo-center) lacks all of the terminal Ediacaran–lowest Cambrian formations in the GSSP area of the southern Burin Peninsula (Stops 4–6B), as well as the entire Bonavista Group (Figures 3, 5). The oldest Cambrian unit is the trilobite-bearing Brigus Formation (Figure 7). The Brigus Formation is thin (31.35 m). Its basal bed (25 cm of neomorphic, pink lime mudstone with limonitic, planar microbial mats, oncoids, and wave-oriented, tube-like *Coleoloides typicalis* conchs) is nonconformable on hydrated Ediacaran volcanics. The lower 10.75 m of the formation are red-brown, purple, and green mudstone alternations with abundant traces but no trilobites (Figure 8). The “FAD” of trilobites is, thus, not a diagnostic local marker for the base of Series 2, Stage 3 (e.g., Peng and Babcock, 2005, 2011), as trilobites, a more offshore biotic element in the Paleozoic (e.g., Westrop *et al.*, 1995), appear in Avalonian upper Cambrian Stage 4 facies only in somewhat more offshore red mudstone with carbonate nodules or fossil-hash tempestites (Landing *et al.*, 1989; Myrow and Landing, 1992; Landing and Westrop, 2004).

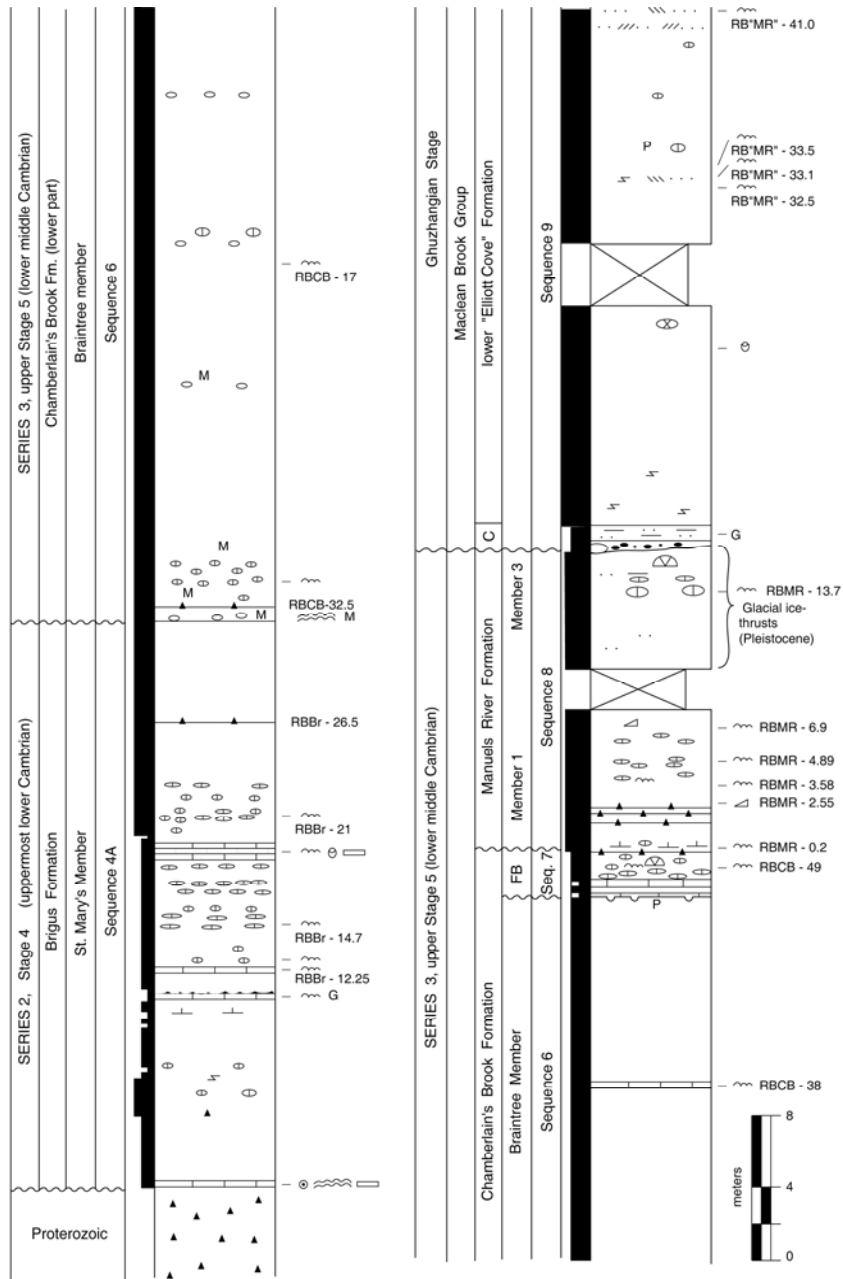


Figure 7. Section through Red Bridge Road quarries, Stop 1. Upper Manuels River Formation and lower McLean Brook Group thicknesses modified by Pleistocene ice-shove thrusts. Fossil ranges in Figure 9, lithologic symbols in Figure 8.

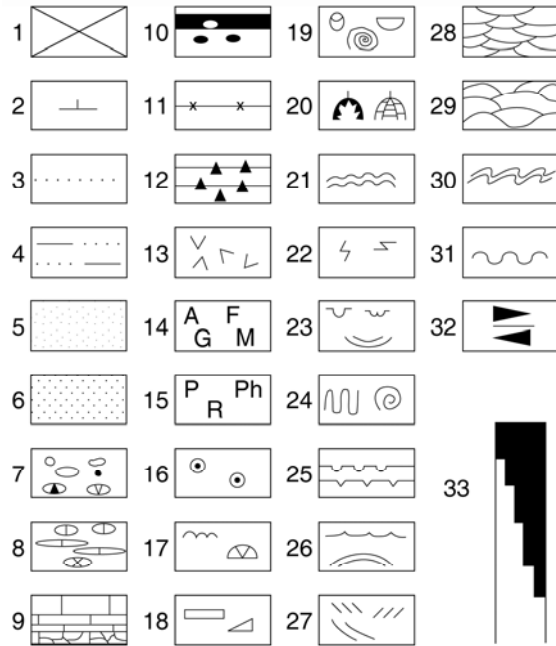


Figure 8. *Lithologic symbols. 1, Covered; 2, siliciclastic mudstone, often slaty, may have calcareous laminae (symbol within rectangle); 3, thin silt- or sandstone bed in mudstone; 4, interbedded thin silt- or sandstone beds in mudstone; 5, thin-medium-bedded sandstone; 6, coarse-grained sandstone; 7, conglomerate with phosphatic (black), volcanic (black triangle), granitic (letter “v”), or mudstone (blank ellipse) clasts; 8, calcareous nodules (upper ellipses), nodular limestones (middle symbol), sideritic nodules (lower symbol); 9, limestone bed (upper) and nodular or small microbial mud mound limestone (lower); 10, phosphatic bed with reworked clasts (upper with ellipse) or nodules (black ellipse); 11, siliceous mudstone with volcanoclastic lenses (middle); 12, volcanic ashes; 13, Ediacaran granite; 14, arkose (A), feldspathic (F), glauconitic (G, often metamorphosed to chlorite), hematitic (H), manganeseiferous (M); 15, pyritic (P), phosphatic (Ph), rhodochrosite (R); 16, oncoids or pisolites; 17, trilobite fragments (upper) and carapaces (lower); 18, conoidal problematica (upper) and hyoliths (lower); 19, brachiopods (upper left), bradoriids (upper right), molluscs (lower); 20, anisograptids (left) and Rhabdinopora (right); 21, stromatolites; 22, Planolites (left), horizontal traces (right); 23, Arenicolites (upper left), arthropod traces (upper right), Teichichnus (lower); 24, Taphrhelminthoida (left), cf. Cyndrichnus; 25, burrow hyporeliefs (upper), desiccation cracks (lower); 26, wave ripples (upper), hummocky cross stratification (lower); 27, microcross-bedded (upper), trough cross-bedded (lower); 28, trough cross-bedded unit; 29, gravel dunes; 30, contorted (slumped) bedding; 31, unconformity; 32, fault shows displacement; 33, rock colors from black (top) and successively downward through medium-dark gray, olive green, red or pinkish, purple-red or purple, and white, buff, or yellow (bottom). Modified from Landing and Westrop (1998a, fig. 6).*

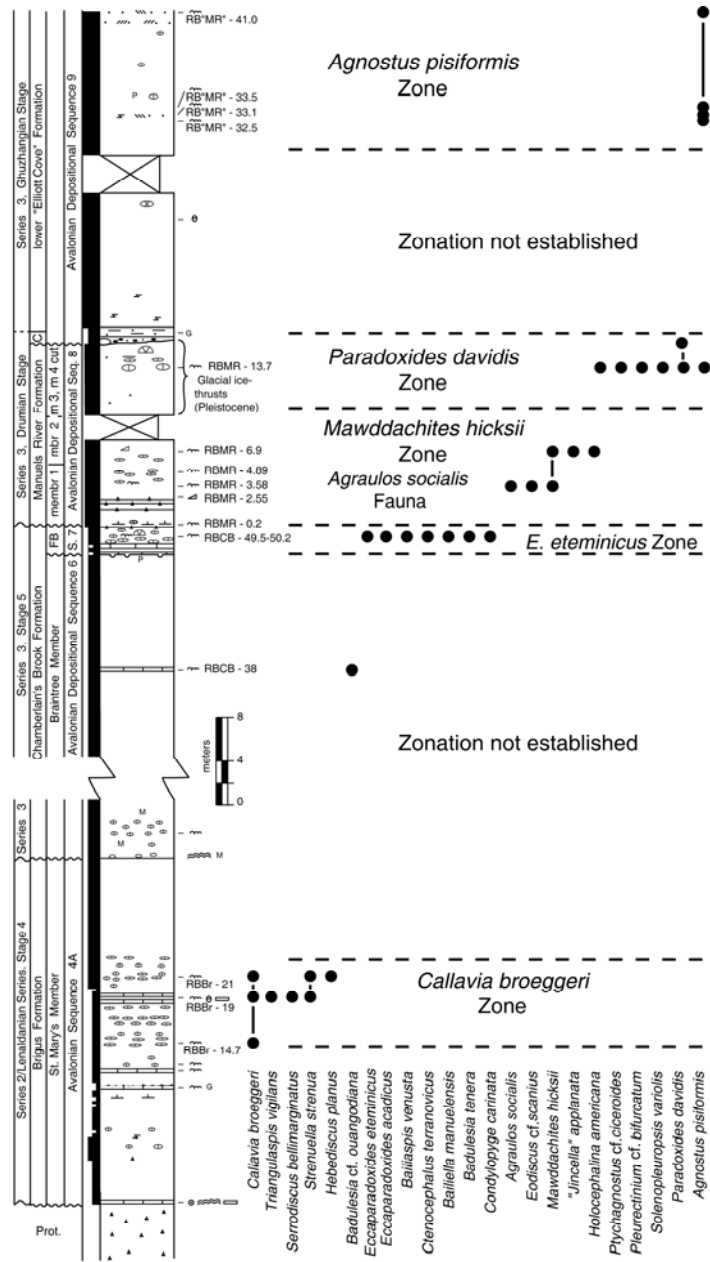


Figure 9. Trilobite and agnostoid arthropod ranges at Red Bridge Road, Stop 1. Detailed section in Figure 7; lithologic symbols in Figure 8. Designation “m 3, m 4 cut” opposite upper Manuels River Formation emphasizes that member 3 is represented and member 4 is cut out by erosion. Modified from Landing and Westrop (1998a, fig. 5); Eccaparadoxides eteminiucus Zone faunas of section in Kim et al. (2002).

A characteristic bed, a variably thick trilobite hash bed, locally glauconitic, with carbonate pebbles and *Myopsostrenua* trilobites at the top of the lower Brigus (top of St. Mary's Member; see Stops 8, 10) is not present at Stop 1. This suggests that the thin Brigus Formation is truncated and referable only to the *Callavia broeggeri* Zone (Figure 9). A 6 cm-thick volcanic ash at 26.5 m (dating in prep.) is likely comparable to the 517.22 ± 0.31 Ma ash from the *Callavia* Zone of the Purley Shale, NW England (Williams *et al.*, 2013).

The base of the Chamberlain's Brook Formation is a cryptic upper lower Cambrian–lower middle Cambrian unconformity within green siliciclastic mudstones. The unconformity is marked by manganiferous carbonate nodules (31.35 m) and manganiferous, laterally linked hemispheroid microbial mats (32 m) in the basal Chamberlain's Brook. As at Manuels River (Howell, 1925), the lower Chamberlain's Brook at Stop 1 is sparsely fossiliferous (inarticulate brachiopods and unidentifiable paradoxidid fragments at 33.5 m and 48 m). It has a volcanic ash at 32.5 m (EL, in preparation). Howell (1925, p. 55) reported *Eccaparadoxides bennetti* from the basal Chamberlain's Brook Formation at nearby Manuels River, and this indicates a substantial unconformity (Avalonian Depositional Sequence 4A–6 hiatus; Figure 5)

Fossils (conocoryphids and paradoxidids of the *Eccaparadoxides etemincus* Zone) are abundant in the uppermost Chamberlain's Brook. They occur in mudstone and limestone beds of the Fossil Brook Member (2.2 m-thick, depositional sequence 7) at the top of the formation (Kim *et al.*, 2002; Figures 5, 9).

The Fossil Brook Member is thicker (3.4 m) at nearby Manuels River (Landing and Westrop, 1998a) and was erosionally truncated before deposition of the Manuels River Formation (Avalonian Depositional Sequence 8; Figure 5) at Red Bridge Road. The Chamberlain's Brook–Manuels River contact is not conformable anywhere in SE Newfoundland

(*vide* Fletcher, 1972, 2003, 2006), as discussed above. A 2 cm-thick, white volcanic ash defines the base of the dysoxic, dark gray–black mudstone-dominated Manuels River Formation, and thin ashes occur higher in the formation in the third quarry (at 1.6, 2.25, 2.45 m). *Hydrocephalus hicksi* Zone trilobites and articulated hyolithids are present through the third quarry (8.4 m).

The upper Manuels River Formation (*Paradoxides davidis* Zone) is present in the fourth quarry. These strata contain numerous Pleistocene ice-thrusts marked by injected sand. A 30–50 cm-thick brown sandstone with phosphatic clasts to 4 cm in diameter caps the Manuels River Formation and marks the Avalonian Depositional Sequence 8–9 contact (Figure 9). As noted above, this thin unit (Cavendish Formation) has the eponymous species *P. davidus* immediately above the Manuels River Formation at Red Bridge Road as well as at Highland Cove in Trinity Bay (Landing and Westrop, 1998a, fig. 10; Figure 2, locality HC). The Cavendish Formation thickens to several meters to the west in Trinity Bay (Westrop and Landing, 1998a) and to 45 m in southern St. Mary’s Bay (Landing and Westrop, 1998a, b; “Beckford Head Formation” of Fletcher, 2003, 2006).

Fossils are not present immediately above the Manuels River Formation (Drumian Stage) and feather-edge of the Cavendish Formation in this section and at Highland Cove, Trinity Bay (Landing and Westrop, 1998b). However, basal Guzhangian Stage (*Lejopyge laevigata* Zone) trilobites occur just above the Cavendish Formation sandstone at Manuels River (Westrop and Landing, 1998a, p. 28). Higher Guzhangian State (*Agnostus pisiformis* Zone) agnostoid arthropods are present 16.7–17.7 m and 25.2 m above the top of the Manuels River in the lower Chesley Drive Group at Red Bridge Road.

Stop 2. Ediacaran succession at Spaniard's Bay (June 23; GMN)

Location

Proceed to the Town of Spaniard's Bay. Follow a secondary road to Green Head. Park at the end of the road and walk along the coastal trail to the headland. DESCEND TO THE ROCK PLATFORM WITH CARE. This stop is most visible at low tide.

Generalized Stratigraphy and Significance

This section shows typical development of the Mistaken Point Formation in the Spaniard's Bay area, and consists mainly of thin- to medium-bedded turbidites (T_{cde} and T_{de}) without any evidence of exposure or shallow-water conditions. The beds are tabular, with sharp basal contacts and normal grading, and are interpreted as representing deposition from weak to moderate, low-concentration turbidity currents in a deep-water setting. Beds and laminae of volcanic ash and tuffaceous sandstone occur sparingly throughout the section. Current ripples in the turbidites indicate that downslope paleocurrent directions are to the east, but oriented fronds in the contourite beds show strong unimodal alignment to the west and southwest (Ichaso *et al.*, 2007). This 90° difference between frond orientations and downslope paleocurrents in turbidites implies that the fronds were oriented by contour currents running along the slope, a motif similar to that in deep-water Ediacaran assemblages in NW Canada (Dalrymple and Narbonne, 1996) and Mistaken Point (Benus, 1988; Wood *et al.*, 2003).

The fossil assemblage is typical of the Mistaken Point Formation, and consists mainly of *Aspidella*, *Fractofusus misrai*, *Charniodiscus spinosus*, and *Bradgatia*. A spectacular surface on the headland contains more than 13 fossils/m², which is one of the highest concentrations of Mistaken Point fossils seen anywhere on the Avalon Peninsula. This fossil assemblage is

compositionally similar to what is seen in the D and E beds at Mistaken Point more than 100 km to the south and to fossil beds that will be observed during this conference on the Bonavista trip 100 km to the north. These observations imply uniform and widespread distribution of the Avalon assemblage.

Stop 3. Ediacaran succession at Upper Island Cove (June 23, GMN)

Location

The locality is near Upper Island Cove on the east side of Conception Bay (Figure 2, location UIC). To limit any unauthorized collecting, we do not provide locality details.

Generalized Stratigraphy and Significance

This locality has been designated a provincial “Significant Paleontological Site.” The collection, marking, or damaging of fossils from this site is prohibited.

The strata at this stop are part of the Trepassey Formation, which overlies the Mistaken Point Formation throughout the Avalon and Bonavista peninsulas. The formation consists of a monotonous succession of thin-bedded, dark grey turbidites (mostly T_{de}) that lack any shallow-water features. Slump beds up to 15 m thick punctuate this succession. Rare current-rippled sandstones show downslope paleocurrents to the south and southeast, which are consistent with the orientation of fossil fronds in the tops of the turbidites. Laminae of volcanic ash occur only sporadically throughout this succession. Deposition probably took place due to the action of weak, low-concentration, turbidity currents on a deep-water slope.

In contrast with Stop 2, the virtual absence of volcanic ash beds precludes Conception Group-style preservation of a diverse fossil assemblage. Consequently, the most common fossils

in the Trepassy Formation at this location and throughout the Avalon Peninsula are *Aspidella* discs that show Fermeuse Formation-style preservation on bed tops beneath an overlying turbidite.

A single bedding plane in this succession exhibits exquisite, locally three-dimensional preservation that is among the best Ediacaran preservation known anywhere and yields critical information about the external and internal microstructure of these fossils (Narbonne, 2004; Narbonne *et al.*, 2007). Spaniard's Bay specimens were molded by mud rather than cast by sand; and this preserves features as small as 30 μm in width on the best specimens (Figure 10A). Information on the sedimentology and taphonomy of this unique Lagerstätten can be found in Ichnaso *et al.*, 2007) and Brasier *et al.* (2013).

All of the fossils in this surface are rangeomorphs, an extinct clade of Ediacaran life that was especially important in older and deeper-water assemblages. The fundamental architectural element was the “rangeomorph frondlet,” a cm-scale structure that exhibits self-similar branching over four fractal scales (Figure 10A, B). These elements were combined as modules to make up the diversity of fronds, bushes, multibranching networks, spindle-shaped recliners, and other forms that dominated the Ediacaran deep-sea floor (Figure 10C–G; Narbonne *et al.*, 2009; Brasier *et al.*, 2012; Dececchi *et al.*, 2017). It is difficult to relate rangeomorphs to any modern group of macroscopic organisms, and they appear to represent a “forgotten” architecture and construction that characterized early stages in the terminal Neoproterozoic evolution of complex multicellular life.

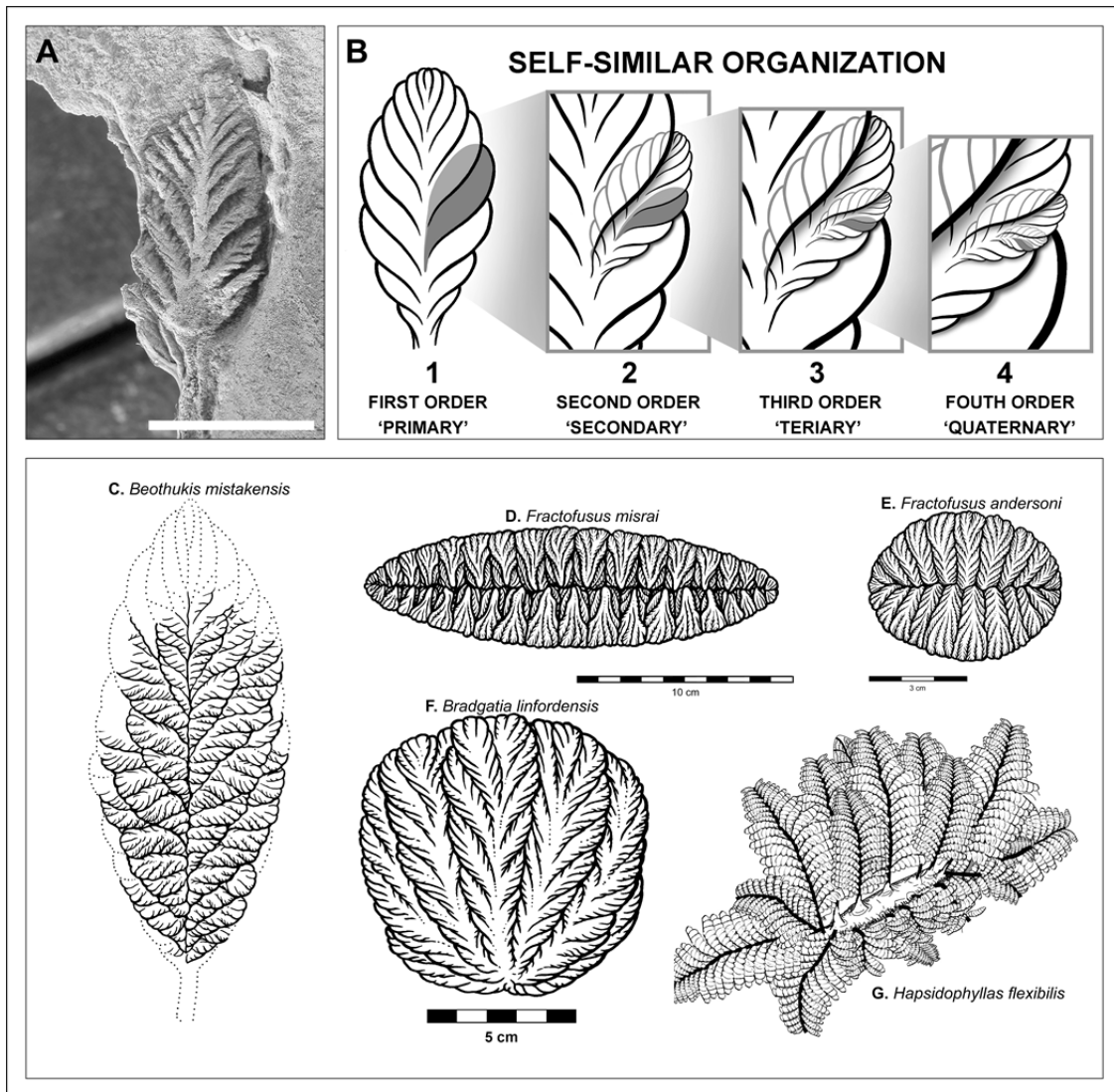


Figure 10. Ediacaran rangeomorphs, Mistaken Point and Trepassy formations, Avalon Peninsula. A, B *Avalofractus abaculus*; A, Holotype from Upper Island Cove (Narbonne, 2004); B, Self-similar branching over four fractal orders (Narbonne et al., 2009); C, *Beothukis mistakensis*, an Ediacaran frond (Narbonne et al., 2009); D, *Fractofusus misrai*, a spindle-shaped organism reclining on the sea floor; E, *Fractofusus andersoni* Gehling and Narbonne, 2007); F, *Bradgatia linfordensis*, bush-shaped organism (Flude et al., 2008); G, *Hapsidophyllas flexibilis*, multibranch network (Bamforth and Narbonne, 2009). Images B–F by Dr. Peter Trusler, Melbourne.

Stop 4. Fortune Head: Ediacaran–Cambrian Boundary GSSP (June 24, PM, GMN, LAB, MGM, BL, RG, EL)

Location

The base of the section (lower Chapel Island Formation, members/lithofacies associations 1–2B) is reached by driving 2.7 km south of Fortune village on Route 220. Turn W on the gravel road, drive across a creek valley and ridge to the lighthouse. Park the vehicles immediately S of the GSSP section (see cover illustration). The section extends west along the coast for approximately 1 km (Figure 11). It is superbly exposed in a continuous series of low cliffs that are readily accessible by foot. The strata dip northwest at 15–46° with dip increasing up-section. A few faults are present, but marker horizons allow for easy correlation across them. The section contains over 400 m of continuous strata (Figure 12), which is broken into stops for convenience.

Stops 4–6B lie in the Fortune Head Ecological Reserve. Specimen collecting or use of hammers and molding compounds without a permit is prohibited under the Newfoundland and Labrador Wilderness and Ecological Reserves Act. Permits for scientific research at Fortune Head or any other ecological reserve may requested from Parks and Natural Areas, Department of Tourism, Culture and Recreation, Government of Newfoundland and Labrador.

Overview of Terminal Ediacaran–Lowest Cambrian Trace Fossil Biostratigraphy and Evolution of the Chapel Island Formation (LAB, MGM, GMN, BL, RG)

During the last few decades there has been a burst of detailed ichnologic studies that access trace fossil successions as part of the Cambrian Evolutionary Radiation (CER) (e.g., Narbonne *et al.*, 1987; Crimes, 1992, 1994; Jensen, 1997; MacNaughton and Narbonne, 1999;

McIlroy and Logan, 1999; Seilacher *et al.*, 2005; Buatois *et al.*, 2014; Mángano and Buatois, 2014, 2016). These studies have focused not only on the importance of trace fossils in biostratigraphy (e.g., Narbonne *et al.*, 1987; Crimes, 1992, 1994; Jensen, 2003; MacNaughton and Narbonne, 1999; MacNaughton, 2007; Mángano *et al.*, 2012), but also on such macroevolutionary aspects as the role of bioturbation in geochemical cycles, ecosystem engineering, and geobiology (e.g., McIlroy and Logan, 1999; Buatois *et al.*, 2014; Mángano and Buatois, 2014, 2016), as well as ichnodiversity and ichnodisparity changes as a response to evolutionary innovations (e.g. Mángano and Buatois, 2014, 2016; Buatois *et al.*, 2016).

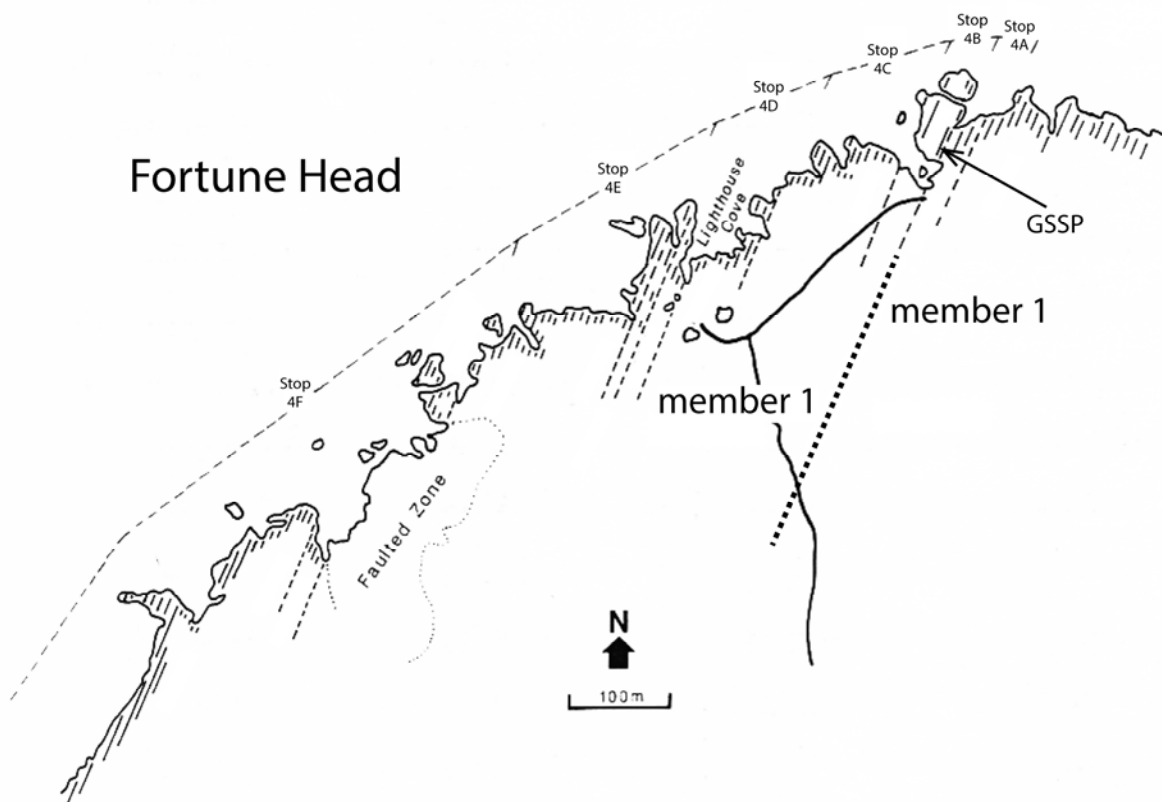
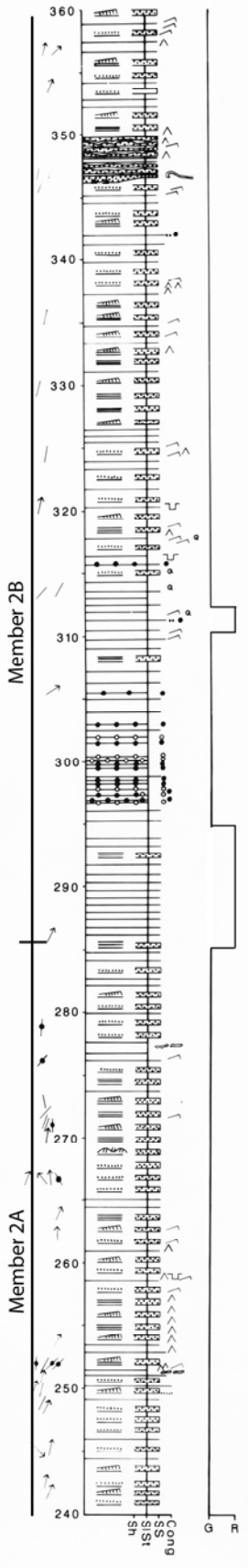
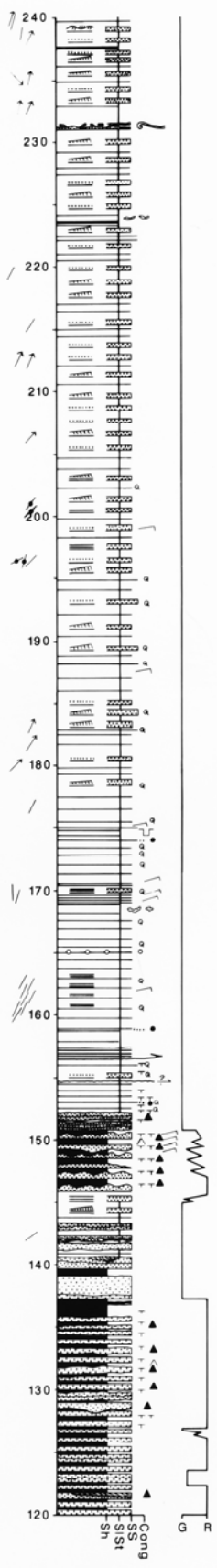
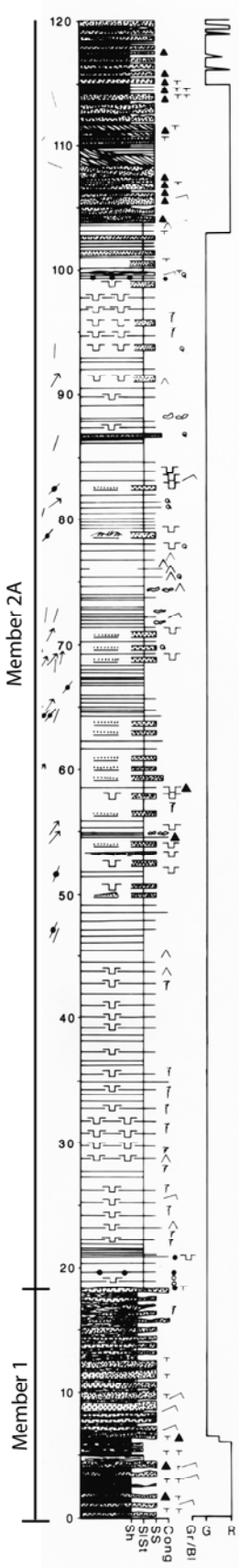


Figure 11. Map of Fortune Head. Dashed lines indicate strike of strata. Only stops 4A–F are shown.



Fortune Head

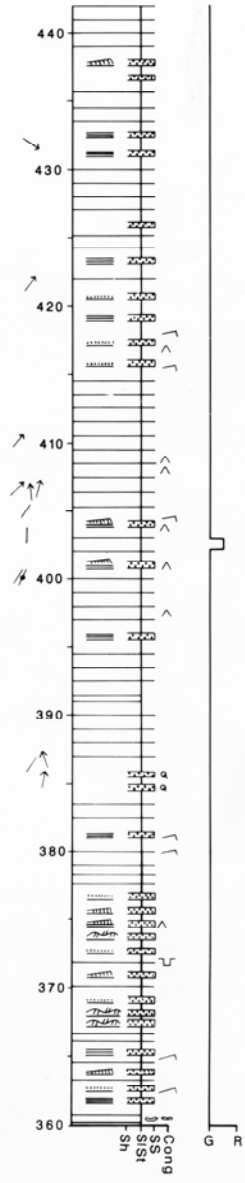
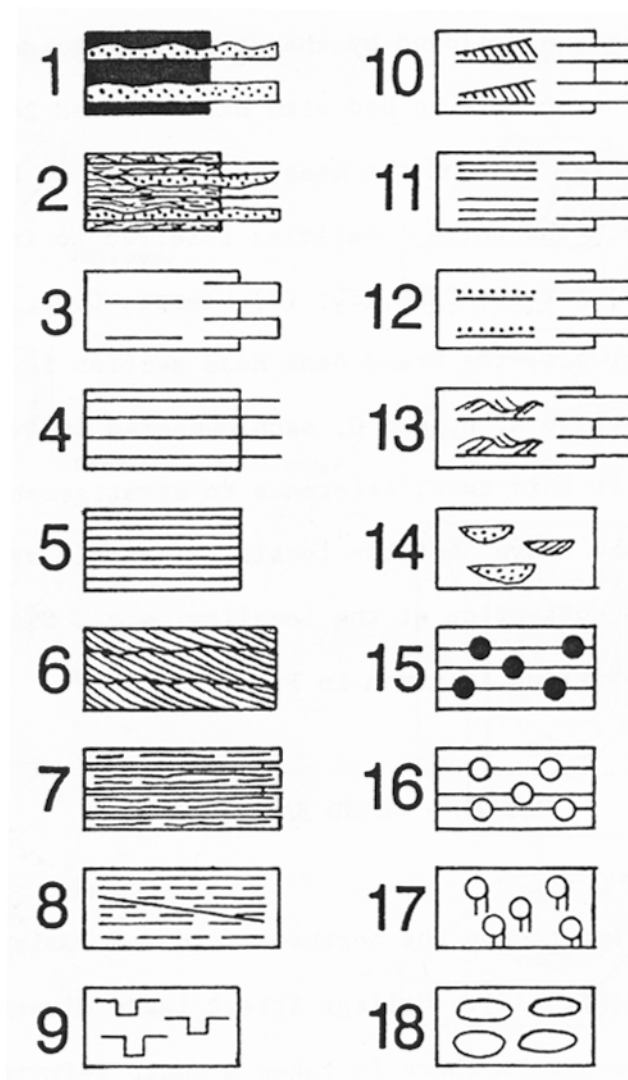


Figure 12. Measured section for Fortune Head. Symbols to left of column are paleocurrents with north at top of page. Color is given by vertical lines to right of column (Gr/Bl, gray/black; G, green; R, red). Informal members 1 and 2A are also lithofacies associations 1 and 2A.



Lithologies: 1, thin-medium bedded, red and green sandstone and shale; 2, gray-blk laminated shale; 3, thin-medium bedded green siltstone and shale; 4, thin sandstone and siltstone; 5, sandstone and laminated siltstone; 6, trough X-bedded sandstone; 7, medium-thick bedded micaceous red ss; 8, as 7 but thick-massive bedded; 9, gutter and pot casts; 10, ripples; 11, parallel lamination; 12, graded; 13, HCS; 14, sandstone scour and fills; 15, phosphate nodules; 16, pyrite nodules; 17, bioturbation; 18, carbonate nodules.

Trace fossils are particularly valuable as sources of information on the CER due to a number of peculiarities of biogenic structures (Mángano and Buatois, 2014, 2016). First, they are typically produced by soft-bodied animals, which tend to have a meager body fossil record. Secondly, they provide information on the presence of a broad spectrum of phyla. Thirdly, their record across the Ediacaran–Cambrian boundary is far more continuous than the body-fossil record. Fourth, as a reflection of animal-substrate interactions, they allow a better integration with sedimentologic data, which leads to finely calibrated interpretations of the associated depositional environments. Fifth, their careful analysis provides evidence of the impact of animal activity on substrate properties, on other organisms, and in geochemical cycling.

There is now general agreement that the Ediacaran–Cambrian boundary is marked by an increase in global and alpha ichnodiversity (Figure 13). The lowest Cambrian (Fortunian Stage) is characterized by a dramatic increase in ichnodisparity, as revealed by the appearance of the main types of categories of architectural design (Figure 14). Furthermore, there is evidence that changes in the number and types of trace fossils were accompanied by an increase in depth and bioturbation through the early Cambrian, as well as by changes in the style of bioturbation (Figure 15).

The 1400-meter-thick succession of fossiliferous, sub-trilobite Cambrian sedimentary strata preserved in the Burin Peninsula represents an outstandingly thick and reasonably continuous record of earliest Cambrian evolution and events. Most of these strata and all of the strata representing the terminal Ediacaran and the earliest stage of the Cambrian consist of fine-grained siliciclastics that were deposited in shallow marine environments, a situation that is typical of most of Avalonia, Baltica, Laurentia, Australia, southern Africa, South America, and other regions removed from the carbonate facies belts of central Asia. The biostratigraphy of

these deposits relies on Ediacaran-type fossils and ichnofossils, supplemented by microfossils and isotope geochemistry where available along with the later appearance and use of shelly fossils. Three successive ichnofossil zones have been recognized in the siliciclastic facies of the Ediacaran–Cambrian Chapel Island Formation in the Burin Peninsula and can be correlated into siliciclastic facies worldwide (Narbonne *et al.*, 1987). A fourth interval, the *Cruziana tenella* Zone (= *C. problematica* Zone) has been tentatively identified in the Random Formation (MacNaughton and Narbonne, 1999). Typical Ediacaran–Cambrian body fossils (Figure 16) and trace fossils (Figure 17) from these zones are recorded in this field guide.

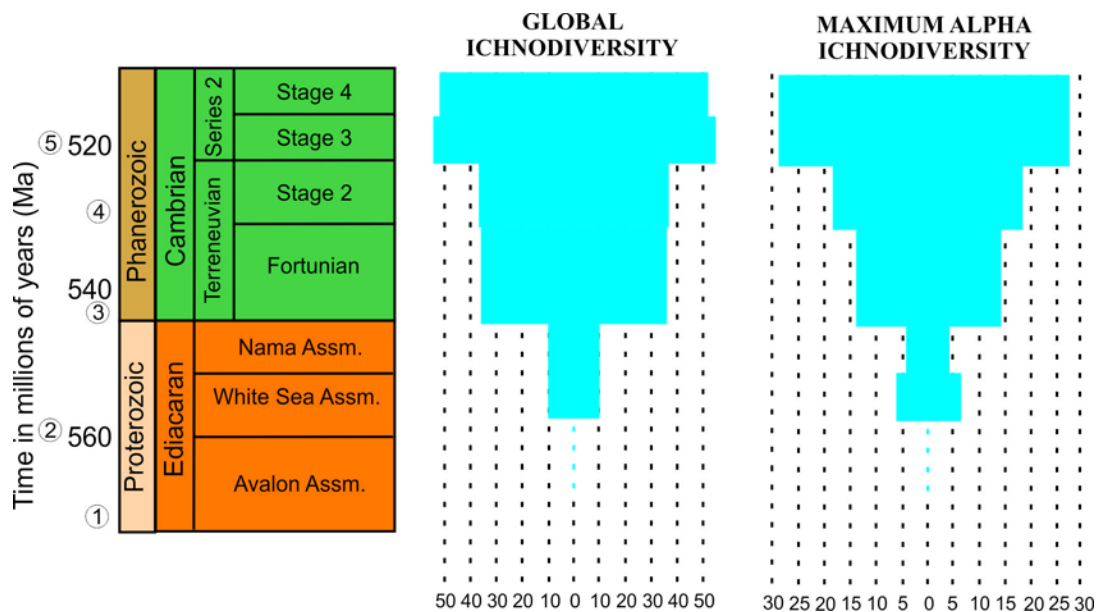
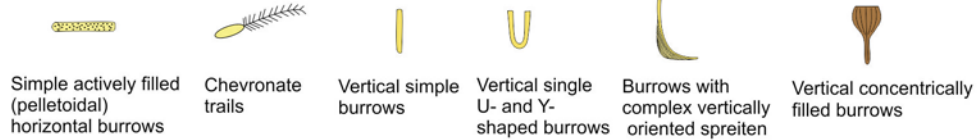


Figure 13. Changes in global and alpha ichnodiversity in the Ediacaran–Cambrian transition (after Mángano and Buatois, 2016). 1, appearance of Ediacaran biota; 2, first uncontroversial evidence of bilaterian trace fossils; 3, major diversification of trace-fossil bauplans; 4, onset of vertical bioturbation and coupling of benthos and plankton; 5, earliest fossil Lagerstätte (Chengjiang) and Cambrian explosion according to body fossils. In contrast to fossil Lagerstätte, the trace-fossil record is continuous through the critical Ediacaran–Cambrian interval. Assm. is assemblage.

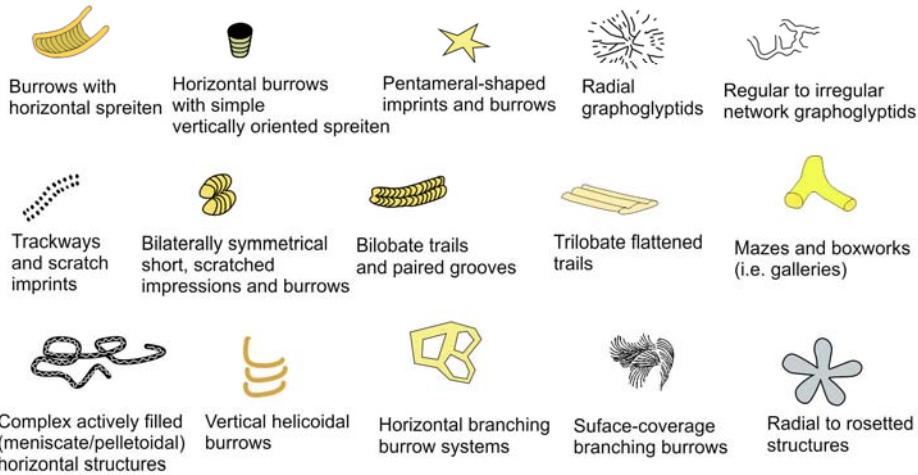
CAMBRIAN STAGES 3 & 4



CAMBRIAN STAGE 2



FORTUNIAN



EDIACARAN

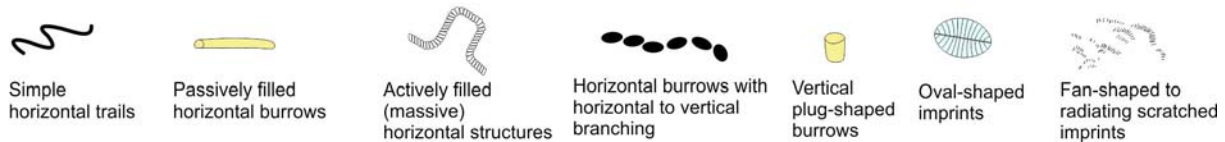


Figure 14. Schematic reconstructions of the main Ediacaran and early Cambrian architectural designs in trace fossils (after Mángano and Buatois, 2016).

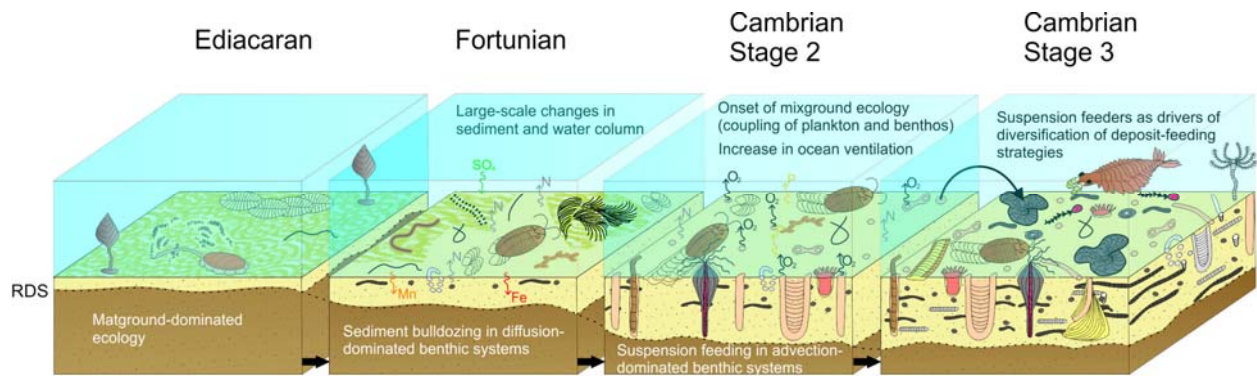


Figure 15. Evolutionary changes in benthic faunas and ecosystem engineering through the Ediacaran–Cambrian transition (after Mángano and Buatois, 2014). Note deepening of the redox discontinuity surface (RDS) and complex feedback loops.

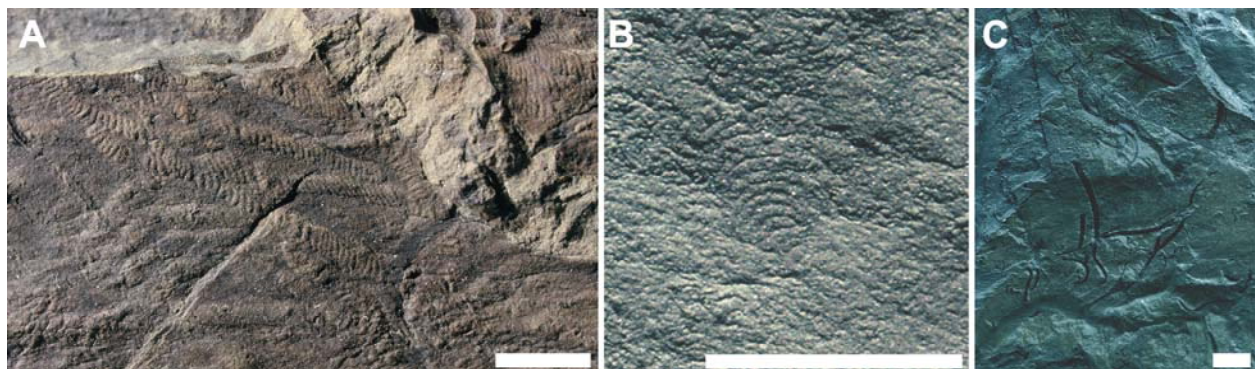


Figure 16. Ediacaran (A, B) and Cambrian (C) body fossils from the lower part of member 2 at Fortune Head. Scale bars represent 1 cm. A: *Harlaniella podolica*, 0.2 m below the GSSP; B: *Palaeopaschichnus delicatus*, 0.2 m below the GSSP; C: *Sabellidites cambriensis*, 165 m above the base of member/lithofacies association 2.

The *Harlaniella podolica* Zone is terminal Ediacaran in age, and brackets member 1 and the basal 2.4 m of member 2 (i.e., lithofacies associations 1–lowest 2A of Landing, 1996a). It is based on the Ediacaran megafossils *Harlaniella podolica* (Figure 16A) and *Palaeopaschichnus*

delicatus (Figure 16B), which were originally regarded as Ediacaran burrows (Palij, 1976; Crimes and Fedonkin 1994), but are now widely regarded as serially repeating Ediacaran megafossils (see reviews in Seilacher *et al.*, 2003; Antcliffe *et al.*, 2011; Ivantsov *et al.*, 2013). *Palaeopascichnus* is strictly restricted to upper Ediacaran strata worldwide (Narbonne *et al.*, 2014) and has been reported from the Ediacaran of Australia (Haines, 2000), Norway (Crimes and McIlroy 1999), Wales (Cope 1983), Ukraine (Palij, 1976), the White Sea (Fedonkin, 1985), China (Dong *et al.*, 2008), Siberia (Grazhdankin *et al.*, 2008), and several stratigraphic levels in Avalonian Newfoundland (Narbonne *et al.*, 1987; Gehling *et al.*, 2000). *Harlaniella* is also globally distributed, with occurrences in the Ediacaran of the White Sea, Ukraine, Australia, and Avalonian Newfoundland (Ivantsov, 2013). Carbonaceous fossils of the annulated tube *Sabellidites* (Figure 16C) and the vendotaenid alga *Tryasotaenia* first appear in this zone and extend upward into Cambrian strata.

Ediacaran trace fossils in the Chapel Island Formation are mainly simple, mm-diameter, subhorizontal, mat-grazing trails such as *Helminthoidichnites*, with first appearance of treptichnids and penetrative burrows in the uppermost few meters of the zone. All of these features are typical of siliciclastic strata of the terminal Ediacaran stage (Xiao *et al.*, 2016).

The ***Treptichnus pedum* Ichnofossil Assemblage Zone** is the basal ichnofossil zone of the Fortunian Stage, the Terreneuvian Series, and the Cambrian System. The designation “*T. pedum* Zone,” with its base defined by the lowest occurrence of *T. pedum* at the future Fortune Head GSSP, was initially proposed for the basal Cambrian ichnofossil zone (Narbonne *et al.*, 1987; Landing, 1992b, 1994; Brasier *et al.*, 1994b). However, as detailed below, *T. pedum* was later found a few meters below the base of the GSSP, and a “*T. pedum* Ichnofossil Assemblage Zone” with its base at the same level as the GSSP but now defined to be within the lower range

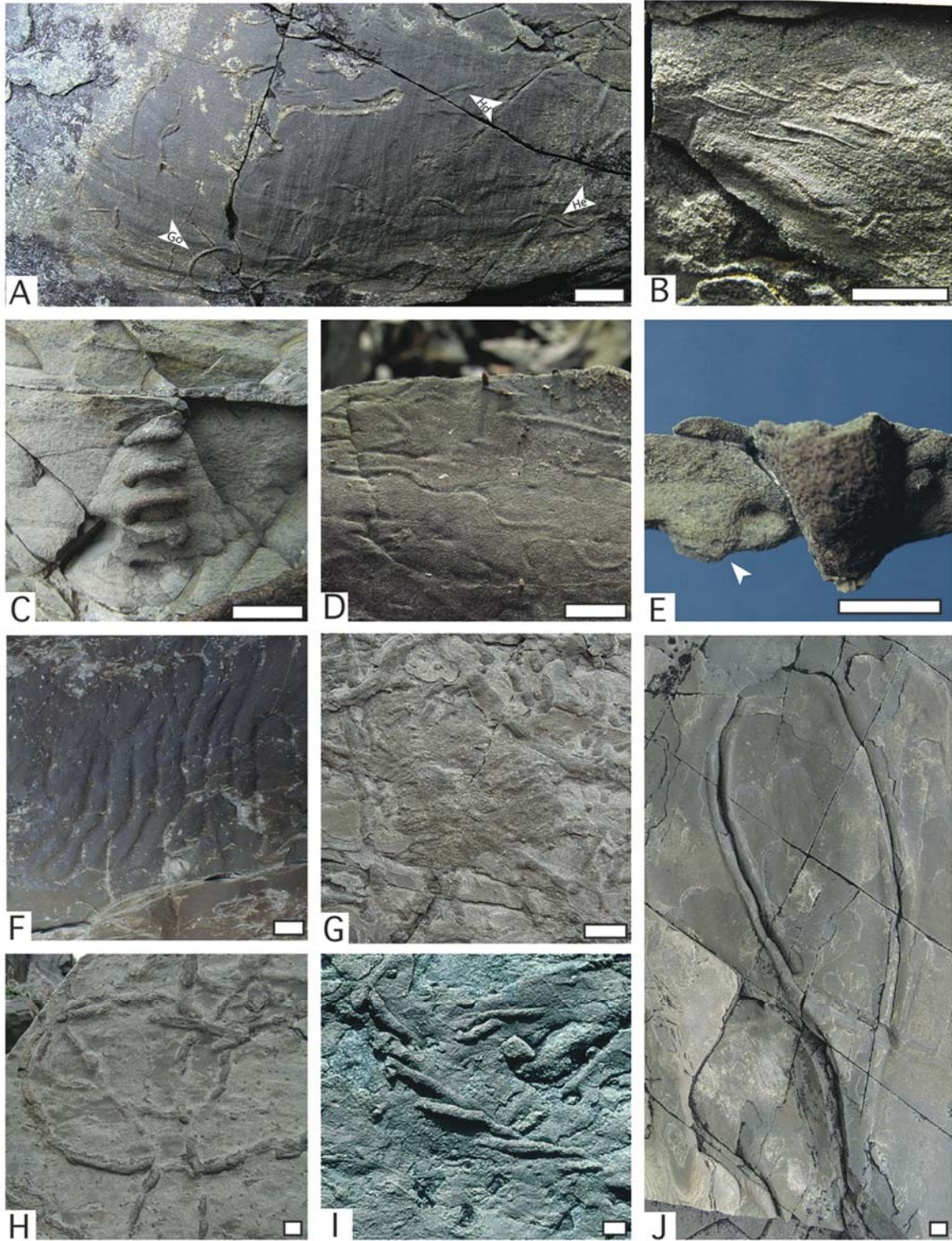


Figure 17. Diagnostic Cambrian ichnofossils of member/lithofacies association 2 of the Chapel Island Formation from Fortune Head and Grand Bank Head. Scale bars represent 1 cm. A, Grazing trails, top view. Go, *Gordia* isp.; Hd, *Helminthoidichnites tenuis*; and He,

Helminthopsis tenuis, Fortune Head. **B**, *Arthropod scratch marks, sole view*. *Monomorphichnus isp.*, Grand Bank Head. **C**, *Spiral burrows, side view*. *Gyrolithes isp.*, Fortune Head. **D**, *Sinusoidal grazing trails, top view*. *Cochlichnus anguineus*, Fortune Head. **E**, *Coelenterate dwelling burrows, side view*. *Conichnus conicus*, arrow points to a small *Bergaueria isp.*, Fortune Head. **F**, *meandering Psammichnites isp.*, Fortune Head. **G**, *Arthropod resting trace, sole view*. *Rusophycus avalonensis*, Grand Bank Head. **H**, *Treptichnus pedum, sole view*, Grand Bank Head. **I**, *Treptichnus pedum, sole view*, Fortune Head. **J**, *Specimen traditionally assigned to Taphrhelminthopsis circularis, but now regarded as a preservational variant of Psammichnites, top view*, Fortune Head.

of *T. pedum* and immediately above the highest occurrences of the characteristic Ediacaran problematica *Harlaniella* and *Palaeopasichnus* was proposed (Landing *et al.*, 2013a; Geyer and Landing, 2016).

There is a marked increase in both trace fossil abundance and diversity, with several notable trace fossil taxa including *Treptichnus* (Figure 17H–I), *Gyrolithes* (Figure 17C), *Conichnus* (Figure 17E), *Cochlichnus* (Figure 17D), and, potentially, *Monomorphichnus* (Figure 17B) first appearing at or near the base of the *Treptichnus pedum* Ichnofossil Assemblage Zone, and this marks a major diversification in the biology, ecology, and behaviour of animals that inhabited the sea floor and substrate (Mángano and Buatois, 2014, 2016; Carbone and Narbonne, 2014). The initial studies of the Fortune Head section showed a burst of originations at the GSSP level (Narbonne *et al.*, 1987) but studies over the subsequent three decades have shown the lowest appearances of these key taxa occur at different levels in the 5 m of strata that bracket the GSSP (Gehling *et al.*, 2001; this study), a finding that is not particularly surprising in view of stratigraphic uncertainties in the lowest appearances of fossil taxa and improved understanding of the significance of confidence intervals in the lowest and highest occurrences of fossil taxa (Strauss and Sadler, 1989; Marshall, 1990; Landing *et al.*, 2013b).

These apparently staggered first appearances of trace fossil taxa through a 5 m interval of rapidly deposited sediments do not diminish the immense biological and geochemical significance of the abrupt increase in the abundance, diversity, and behaviour of burrowing organisms at the base of the Cambrian (Callow and Brasier, 2009; Mángano and Buatois, 2014). They do, however, call for increased caution and investigation of the biological and geochemical criteria used in recognizing and correlating the basal Cambrian GSSP. The ichnotaxa changes in the uppermost Ediacaran at the Fortune Head GSSP reflect global changes in increased diversity, including the appearance of trichophycids in the terminal Ediacaran. Particularly encouraging for the continued utility of the GSSP at Fortune Head is the surprisingly eurytopic nature of *T. pedum*, its likely rapid distribution globally after the form's evolution, and the presence of a strong carbon isotope excursion in the terminal Ediacaran and below the lowest occurrence of *T. pedum* in Oman and South China (e.g., see review in Landing *et al.*, 2013b). The details of trace fossil distribution in the Fortune Head section are part of an ongoing study by B. Laing (University of Saskatchewan), and her findings to date are summarized in Figure 18.

The ***Rusophycus avalonensis* Zone** begins approximately 130 m above the base of member/lithofacies association 2 of the Chapel Island Formation and continues to at least the top of the Chapel Island Formation (Fortunian through Cambrian Stage 2). The base of the *R. avalonensis* Zone is marked by the appearance of the oldest trilobite-type resting burrows (*R. avalonensis*; Figures 17G, 19E), along with numerous trilobite-type scratch marks (e.g., *Allocotichnus* [Figure 19A, C, D] and *Dimorphichnus*). Other key trace fossil taxa that also have a lowest appearance in the *R. avalonensis* Zone include the vertical U-shaped spreiten burrow *Diplocraterion*, the large meandering trail *Psammichnites* (Figure 17F, J), the spreiten feeding trace *Teichichnus rectus*, the trilobite trail *Cruziana fasciculata*, the plug-shaped burrow *Astropolichnus*, the bilobate trail *Didymaulichnus miettensis*, and the branching burrow *Phycodes palmatum* (Mángano and Buatois, 2016).

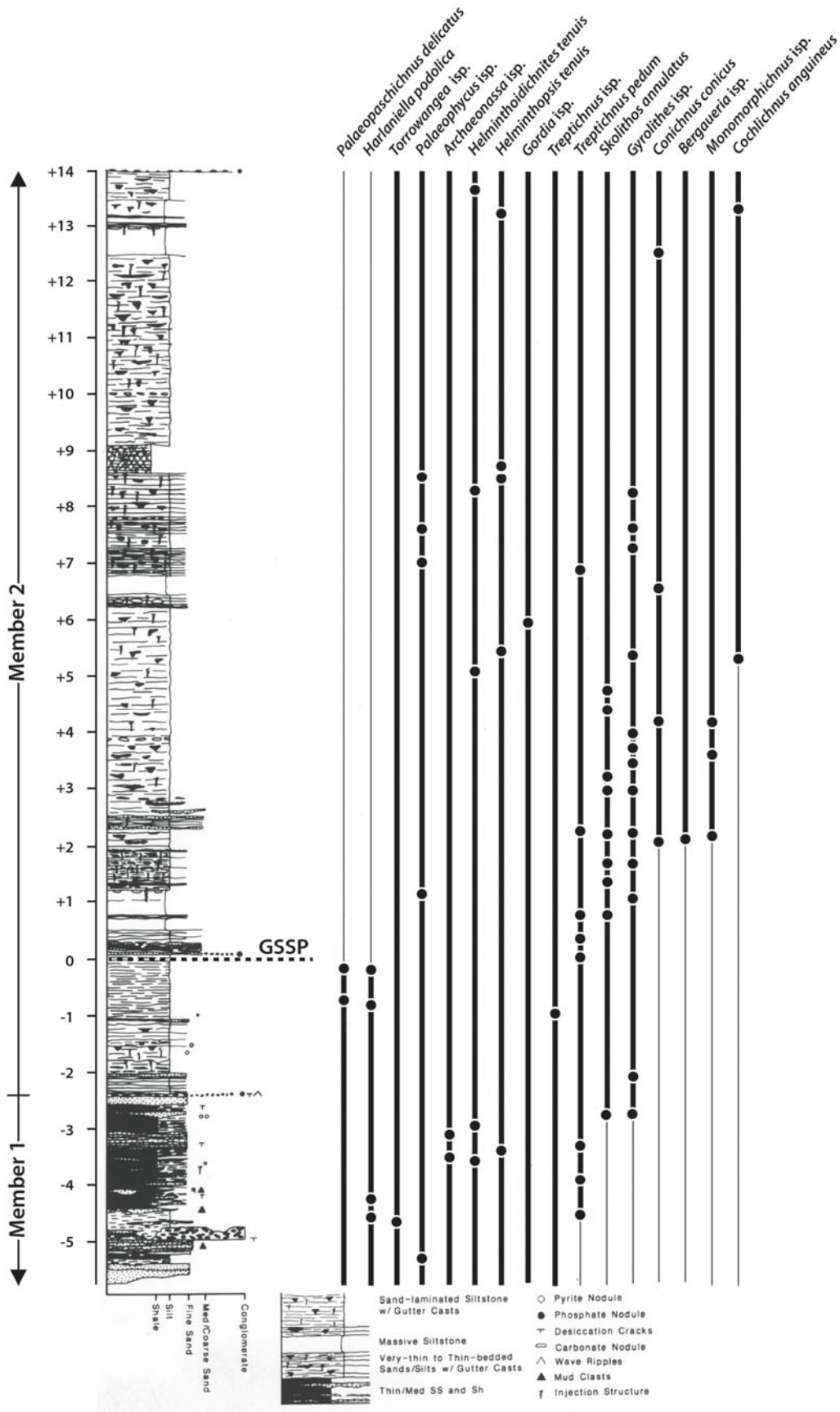


Figure 18. *Fossils across member/lithofacies association 1–2 boundary at Fortune Head. Based on field data collected by Brittany Laing and coworkers (unpublished data). Additional information from Narbonne et al. (1987) and Gehling et al. (2001). Scale in meters.*

This zone is characterized by a significant increase in ichnodiversity in comparison with the *Treptichnus pedum* zone (Mángano and Buatois, 2014, 2016). Ichnologic studies worldwide indicate that the *T. pedum* and *R. avalonensis* zones are recognized globally, including the Mackenzie Mountains of Canada (MacNaughton and Narbonne, 1999), the Flinders Ranges of Australia (Jensen *et al.*, 1998; Droser *et al.*, 1999), the Olenek Uplift of Russia (Rogov *et al.*, 2015), the Yangtze Platform of China (Crimes and Jiang, 1986; Zhu, 1997), and the Alborz Mountains of Iran (Shahkarami *et al.*, in press a, b).

The appearance of *Rusophycus* low in the Fortunian Stage of the lower Cambrian is particularly significant from an evolutionary perspective because these trace fossils are generally regarded as having been produced by arthropod-grade organisms with bilateral symmetry and segmented limbs, an apomorphy of the Arthropoda (Budd and Jensen, 2000). *Rusophycus* specimens from member/lithofacies association 2 of the Chapel Island Formation have been regarded as the oldest unequivocal evidence for the existence of crown-group Ecdysozoa, Lobopodia, and Arthropoda in biotic evolution (Benton *et al.*, 2015; Wolfe *et al.*, 2016). This determination has important implications for our understanding of Cambrian stratigraphy and evolution, but also brings a cautionary note to Cambrian fieldwork. *Rusophycus*, *Dimorphichnus*, *Diplichnites*, and *Allochotichnus* in the *Rusophycus avalonensis* Zone provide reasonably unequivocal evidence that they were constructed by arthropods, but occurrences of *Monomorphichnus* need to be carefully assessed because this ichnotaxon can be difficult to distinguish from inorganic sedimentary structures such as brush marks. This caution should be kept in mind as we search for the oldest unequivocal evidence for arthropods during this field trip in the Chapel Island Formation and in subsequent studies worldwide.

Ediacaran and Early Cambrian Substrates (LAB, GMN, MGM, BL, RG)

Ediacaran substrates in the shallow marine realm worldwide were pervasively covered by microbial coatings that effectively sealed the sediment surface (Gehling, 1999; Seilacher, 1999). There is scant evidence for deep burrowing or bioturbation anywhere in the Ediacaran, and most Ediacaran burrowers were mm-diameter animals that tunneled horizontally through the microbial mat as they mined it for food. These animals produced small, exclusively horizontal trace fossils that typically lack either true branching or systematic meanders (Carbone and Narbonne, 2014; Mángano and Buatois, 2014; Buatois and Mángano, 2012, 2016). These mats also acted as a geochemical seal that separated oxygenated bottom waters in the world's oceans from highly euxinic pore fluids in the sea bottom immediately beneath the microbial mats (Seilacher, 1999; Callow and Brasier, 2009).

Microbially induced sedimentary structures (MISS) and associated trace fossils are common throughout member 2 (Figure 19) and occur through the *Treptichnus pedum* and the *Rusophycus avalonensis* zones (Buatois *et al.*, 2014). In particular, wrinkle marks are common, whereas small gas domes are present locally. MISS typically occur at the top of very fine-grained silty sandstone and are covered by a thin veneer of siltstone. Some of the best exposures to observe the association of trace fossils and MISS are located west of the Lighthouse.

Typical trace fossils associated with MISS include grazing trails (e.g., *Cochlichnus anguineus*, *Helminthopsis tenuis*, *Helminthoidichnites tenuis*), which record grazing of organic matter concentrated within microbial mats below a thin veneer of sediment (Buatois *et al.*, 2014). Other trace fossils associated with MISS in these strata are arthropod trackways, scratch marks, and resting traces, namely *Allocotichnus dyeri*, *Diplichnites* sp., and *R. avalonensis*. Preservation of tiny scratch marks can show delicate morphologic details on bedding tops, and

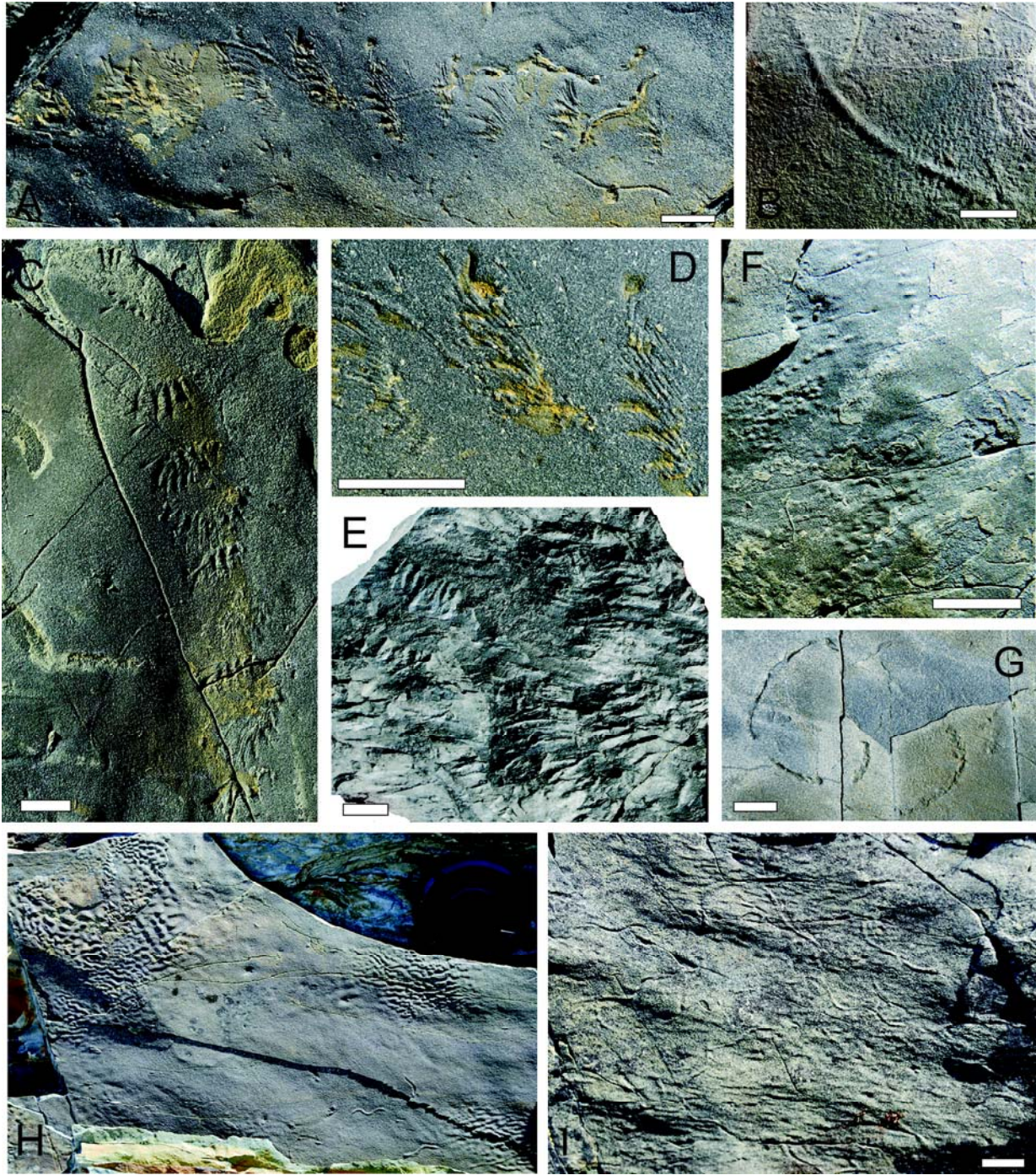


Figure 19. *Microbially induced sedimentary structures and trace fossils in member/lithofacies association 2 of the Chapel Island Formation. Images of upper bedding surfaces. Scale bar is 1 cm. A, Allocotichnus dyeri, note associated Helminthoidichnites tenuis. B, Close-up of Helminthoidichnites tenuis associated with wrinkle marks. C, Allocotichnus dyeri., D, Close up of Allocotichnus dyeri in C, shows fine morphologic details. E, Negative epirelief of trilobite-like*

resting trace Rusophycus avalonensis. *F*, Small gas domes and tiny Helminthoidichnites tenuis. *G*, Treptichnus coronatum. *H*, Wrinkle marks and Cochlichnus anguineus. *Lens cap diameter 5.5 cm*. *I*, Poorly preserved wrinkle marks covered with grazing trails of Cochlichnus anguineus and Helminthopsis tenuis.

also implies significant substrate consistency as would be expected in a rigid microbial mat (Buatois *et al.*, 2014). Finally, the third type of trace fossils present in these deposits are three-dimensional branching burrows, such as *T. pedum*, *T. coronatum* and *T. pollardi* (Buatois *et al.*, 2014).

The presence of matgrounds and associated trace fossils in these outcrops shows that the quintessential Ediacaran-style ecology persisted into the Fortunian. However, the evolutionary innovations of the CER were conducive to more complex interactions between animals and matgrounds. The unique combination of a microbial matground-based ecology and the appearance of new body plans and locomotory mechanisms resulted in a diversity peak of animal-matground interactions during the Fortunian (Buatois *et al.*, 2014; Mángano and Buatois, 2016).

As noted above, Fortunian deposits are characterized by the extensive development of firm substrates close to or at the sediment-water interface (Droser *et al.*, 2002). Trace fossil evidence of firmgrounds in these deposits include: (1) preferential preservation of minute epifaunal and very shallow-tier infaunal trace fossils, (2) scarcity or absence of mid- and deep-tier trace fossils, (3) absence of mottled bioturbation textures, (4) presence of open burrows in mudstone without any wall reinforcement, and (5) unusual styles of burrow preservation, such as adhering and floating.

Geochemical Implications (AJK, GMN)

The gradual decline of Ediacaran-style matgrounds through the Fortunian is also evident in the geochemical record of these strata. High-resolution geochemical analyses in the Chapel Island Formation at Fortune Head and Little Dantzic Cove (Kaufman *et al.*, 2017) show a notable positive $\delta^{13}\text{C}$ excursion in organic matter that starts immediately above the Ediacaran–Cambrian GSSP and then returns to baseline values immediately before the first appearance of biomineralized shells. Similarly, sulfur isotope compositions of pyrite exhibit significant ^{34}S depletion within meters of the Ediacaran–Cambrian boundary at Fortune Head. These isotope anomalies likely reflect the oxidation of reduced carbon and sulfur compounds in sediments as a consequence of matground penetration. Overall, the isotopic shifts in the Chapel Island Formation reflect rapid changes in carbon and sulfur cycling immediately above the Ediacaran–Cambrian GSSP, with stabilization of the sedimentary system shortly before the first appearance of biomineralized fossils in Newfoundland.

Stop 4A. (June 25; PM, GMN, EL)

Location and Lithology

The outcrop is reached by descending a small gully on the east side of a promontory located just east of a small cove. The gently dipping redbeds may be difficult to reach at high tide. Black shale and gray silty shale form the uppermost part of member/lithofacies association 1. These are cut by a minor fault in the notch of the cove. The top of member/lithofacies association 1 is located on the prominent bench of the promontory. The strata are slippery when wet, and participants should exercise caution in examining some parts of the outcrop.

Upper member/lithofacies association 1 (Figure 12) contains abundant features indicative of shallow marine facies, including abundant synaeresis cracks, desiccation cracks, and current ripples. The conspicuous bench marking the member/lithofacies association 1–2 boundary has a number of interesting features including desiccation cracks and abundant phosphatic shale clasts. At the end of the promontory, there is a shallow elongate scour that is covered with phosphatic shale clasts up to 15 cm in length. Wave ripples on the bedding surface above the scour continue down into the scour.

Myrow and Hiscott (1993) interpreted the redbeds of member/lithofacies association 1 to represent peritidal deposits, including possible tidal flat facies. Gray to black shale and silty shale facies were interpreted to record semirestricted shoreline environments, with possibilities including partially enclosed shoreline embayments, estuaries, or interdistributary bays. The surface that represents the member/lithofacies association 1–2 boundary has been interpreted as a minor disconformity surface that shows evidence of subaerial exposure and reworking by waves after submergence (Myrow and Hiscott, 1993). It is certainly a marine flooding surface that records deepening into subtidal siltstone-dominated strata (Stop 4B).

Paleontology

Overall, trace fossils are rare in these Ediacaran strata, and the simple grazing trail *Helminthoidichnites tenuis* is the dominant form (Narbonne *et al.*, 1987). Simple specimens of *Treptichnus* isp. and well-preserved examples of the carbonaceous small shelly fossil *Sabellidites cambriensis* and the vendotaenid alga *Tyrasotaenia* occur in shaley beds in the uppermost few meters of member 1 (Narbonne *et al.*, 1987; Gehling *et al.*, 2001). The lowest known specimen of *Treptichnus pedum* occurs approximately 2 m below the top of member/lithofacies association

1 (4 m below the GSSP; along with the lowest known specimens of *Gyrolithes* sp. and burrows attributed to *Skolithos annulatus*; see Gehling *et al.*, 2001).

Stop 4B. Lowermost Member/Lithofacies Association 2 at the Cambrian Global Section and Stratotype Point (GSSP) (PM, GMN, EL)

Description

The Precambrian–Cambrian Boundary Stratotype Point is located 2.4 m above the base of member/lithofacies association 2 (Figures 11, 12). The strata exposed on this promontory, described as a “gutter cast facies” (Myrow, 1987; Myrow, 1992a, b; Myrow and Hiscott, 1993), are composed of laminae to very thin beds of green-gray sandstone and silver-green siltstone, with common to abundant gutter casts. The sandstone content, excluding gutter casts, varies roughly from 10–40%. White-weathering, quartz sandstone beds make up about 5% of this facies. Sandstone beds greater than a few centimeters are rare. Ptygmatically folded sandstone dikes are abundant, and wave ripples are locally present on thin sandstone beds. Gravity-flow deposits (Figure 20) make up ~10–15% of the facies (Myrow and Hiscott, 1991). These include structureless (massive) and ‘raft-bearing’ siltstone beds a few tens of cm thick, most with lenticular, shallow, channel-like geometries. The raft-bearing beds are siltstone layers with large clasts of interlaminated sandstone and shale that were derived from failure of the facies with which they are interbedded. Carbonate nodules are present within this facies, and in some cases an entire bed experienced early carbonate cementation.

The gutter casts are erosional structures that range from downward-bulging sole structures to isolated channels. The gutter casts at this stop are composed of fine sandstone, and less commonly of coarser sandstone and granule to pebble conglomerate. The sandstone gutter

casts are massive, planar-laminated, or contain oscillatory-flow-type laminae. They have a wide variety of shapes in both in cross-section (Figure 21) and plan view. Sole markings on the sides and bases of these beds include groove marks, poorly developed prod and flute marks, and postdepositional trace fossils. Groove marks are generally parallel to subparallel to the long axes of the gutter casts and form circular to downward spirals on pot casts (Myrow, 1992b). The long axes of the gutter casts show a strong northeast–southwest orientation. The pot casts range in shape from discs to rounded loaflike forms to tall pillars, and are, in some cases, remarkably small (1 cm in diameter) to very large (nearly 20 cm in diameter). They commonly have a corkscrew shape similar to potholes found in bedrock along modern rivers.

Interpretation

The abundant erosional structures (pot and gutter casts) and the thin nature of the sandstone beds in this facies, were used to construct a bypass facies model for tempestite deposition along a fine-grained shoreline (Figure 22) (Myrow, 1992a). This model differs from the standard models in that the most proximal shallow subtidal setting (gutter cast facies) is primarily a zone of bypass, an area dominated by erosion and throughput of sediment. A proximal zone of bypass requires that depositing flows achieve sufficient momentum for bypass and erosion very close to shore (Myrow *et al.*, 1988). Myrow (1992a) proposed that the facies records hyperpycnal flows that were delivered directly from distributary channels. Many storm-influenced deposits have been interpreted as hyperpycnites in the last 15 years (Myrow *et al.*, 2002, 2008; Lamb *et al.*, 2008; Zavala and Arcuri, 2016), and these Chapel Island strata have characteristics consistent with this interpretation. Facies that record the downstream evolution of these flows (i.e., prodelta and shelf) make up the overlying strata.

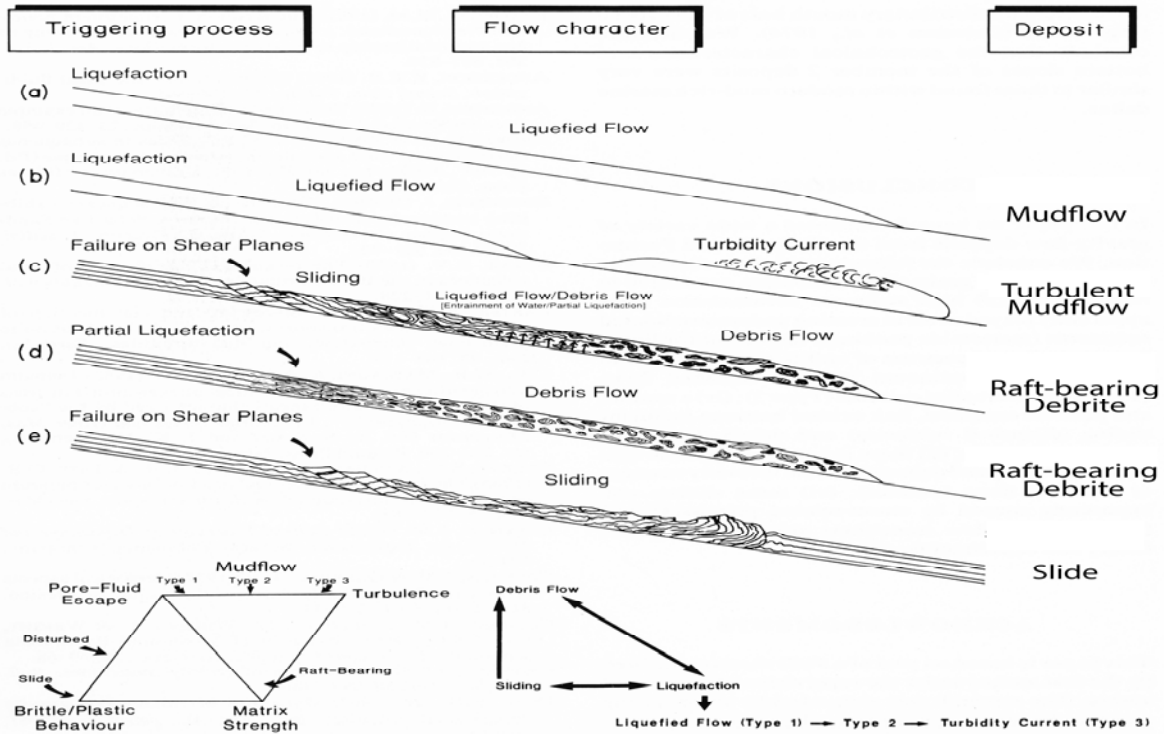


Figure 20. Possible flow transitions (a–e) for the mass flow deposit of member/lithofacies association 2. Flow transitions are summarized with arrows in the lower right part of the figure. Quadrilateral in lower left of figure shows interpreted support mechanisms for the different bed types at time of deposition. Modified from Myrow and Hiscott (1991, fig. 19).

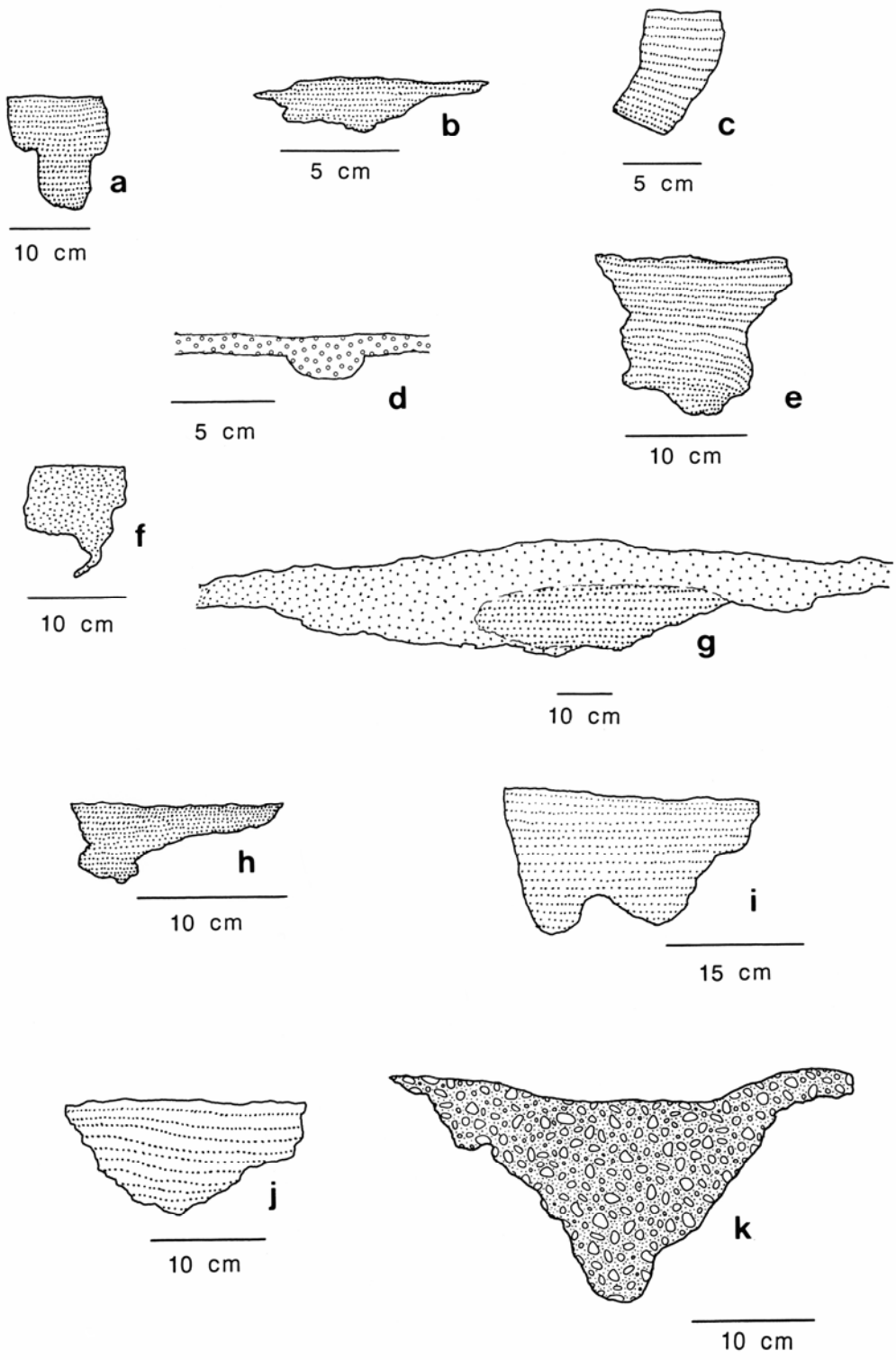


Figure 21. Cross-sectional views of gutter casts of the gutter cast facies. Modified from Myrow (1992b, fig. 4).

Bypass Facies Models

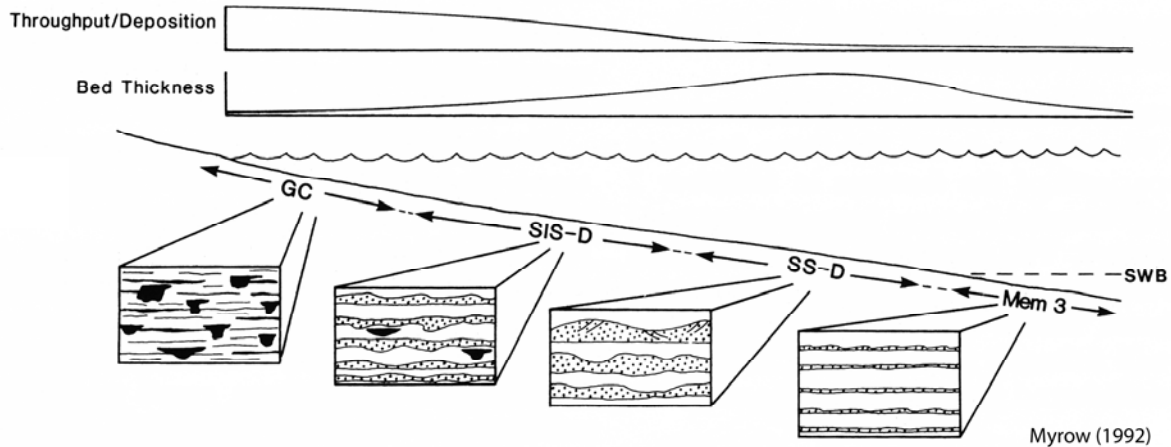


Figure 22. *Tempestite facies model for member/lithofacies association 2 showing increasing then decreasing bed thickness away from shore. The proximal setting is one of bypass of currents where erosion is dominant and gutter and pot casts are abundant. Seaward, gutter cast die out, and hummocky cross-stratification is present in more offshore position. Modified from Myrow (1992a, fig. 16).*

The gutter and pot casts are current erosion structures cut by bypassing flows. They were cut and rapidly filled, and wave ripples on the top of some gutter casts formed during the late stages of deposition. The pot casts were cut by vertically oriented eddies into cohesive mud that used sand as an abrasive, without the need of a large tool (common in bedrock potholes) (Myrow, 1992b).

The gravity flow bed deposits include debris flows (raft-bearing beds) and liquefied mud flows (massive siltstone) (Figure 20). The silt abundance in these beds predisposed them to liquefaction and downslope flow (Myrow and Hiscott, 1991). Given their characteristics, including their channelized nature, they were likely deposited in an upper muddy delta-front setting with failure of the local sea floor or muddy delta-front mouth bars. Prior *et al.* (1986) described silty flows from the Huanghe (Yellow River) Delta that are excellent analogs, and

include silt-flow gullies that are filled with acoustically transparent sediment. The suite of gravity flow deposits, lack of amalgamated sandstone beds, and abundant sedimentary dikes are consistent with high sedimentation rates, high pore pressures, and deltaic deposition.

Paleontology

The Global Standard Stratotype-section and Point (GSSP) for the Ediacaran-Cambrian boundary (Narbonne *et al.*, 1988; Landing, 1992b, 1994; Brasier *et al.*, 1994b) was originally proposed at the base of the *Treptichnus pedum* Zone, 2.4 m above the base of member/lithofacies association 2 (Figure 18). Uppermost Ediacaran strata in the basal 2.4 m of member 2 have a low-diversity assemblage with the same ichnotaxa as uppermost member 1. Significantly, the GSSP is only 0.2 m above the last occurrence of *Palaeopascichnus* and *Harlaniella* (Figure 16A, B), two cosmopolitan taxa widely regarded as serially repeating Ediacaran megafossils (Seilacher *et al.*, 2003; Antcliffe *et al.*, 2011, Ivantsov *et al.*, 2013; see also “Biostratigraphy and Evolution,” above). As the lowest known specimens of *T. pedum* occur several meters below the base of the original “*T. pedum* Zone,” Landing *et al.* (2013b) proposed that the GSSP should be defined (at the originally proposed level) low in the *T. pedum* range and at the top of the ranges of the Ediacaran taxa *Palaeopascichnus* and *Harlaniella*. This horizon is the base of the *T. pedum* Ichnofossil Assemblage Zone.

In addition to *Treptichnus pedum*, basal Fortunian strata show vertical bioturbation as indicated by the corkscrew-shaped burrow *Gyrolithes* *isp.* (Figure 17C) and burrows attributed to *Skolithos annulatus* (Gehling *et al.*, 2001). The morphology of these trace fossils, such as the presence of open burrows in mudstone without wall reinforcement, unusually high quality of preservation (particularly of minute structures), and absence of well-developed mottled textures

suggest firm substrates close to or at the sediment-water interface (Droser *et al.* 2002; Tarhan and Droser, 2014). These strata illustrate the typical components of the *Treptichnus pedum* Ichnofossil Assemblage Zone (Landing *et al.*, 2013b; Laing *et al.*, 2016). The lowest specimens of the probable anemone burrow *Conichnus* (Figure 17E) and the lowest reported possible arthropod scratch marks (*Monomorphichnus*) also occur low in the *T. pedum* Ichnofossil Assemblage Zone (Narbonne *et al.*, 1987). These strata also host the locally lowest convincing examples of *Cochlichnus anguineus* (Figure 17D).

Stop 4C. Coastline from Stop 4B to Red Bed Interval (45–103.7 m)

Description

The lower part of this stop exposes the gutter cast facies and overlying sandstone–siltstone facies, which makes up the bulk of member/lithofacies association 2. This facies has thicker sandstone beds, very few pot and gutter cases, and very abundant raft-bearing beds (debrites).

The sandstone beds show grading and stratification that includes parallel lamination and oscillatory- and combined-flow lamination (i.e., draping and offshooting laminae, form-discordant ripples). Sole marks are well developed and show northeast-directed paleocurrents consistent with those from across Fortune Bay (Figure 23). Vertical changes in stratification indicate that the early stages of deposition were dominated by strong unidirectional flow, and the late stages by oscillatory flow. There are many medium- to very coarse-grained, white-weathering, carbonate-cemented, quartz arenite beds that are dominantly thin, discontinuous, and contain wave ripples. The strata directly below the overlying redbeds are another interval of the gutter cast facies with a few scattered syneresis cracks and pyrite nodules.

Interpretation

This stop shows a deepening and shallowing cycle from nearshore deposits of the gutter cast facies to the slightly deeper-water sandstone–siltstone facies and back into the gutter cast facies. The shift to thicker, more extensive sandstone beds and fewer gutter casts is interpreted by increased distality into a zone of greater deposition and less erosion (Figure 22). This shift also includes a change from abundant massive siltstones (mudflow deposits) beds to raft-bearing debrites. The quartz sandstones are not only better sorted but also coarser grained than most of the other tempestite deposits. This suggests, along with abundant of wave ripples and wave ripple lamination, that these beds were formed by shoreline erosion of local quartz sand sources and transport into the shallow subtidal.

Fossils

Trace fossils are abundant and occur on virtually every sole. Most are simple non-diagnostic forms, but *Treptichnus pedum* and other typical ichnofossils of the *T. pedum* Ichnofossil Assemblage Zone occur sporadically.

Stop 4D. Thick Sandy Red Bed Interval (104–137 m)

Description

The lower 15 m of this 33 m-thick unit of red sandstone and shale are exposed on low cliffs east of “Lighthouse Cove” (Figure 24). They consist of red, thin- to thick-bedded, massive and laminated, well-sorted sandstone, siltstone, and shale. The sandstones have erosional lower and upper surfaces and are mostly parallel-laminated, although a few have trough cross-bedding. Overall, the unit is a fining-up succession with abundant syneresis and desiccation cracks, small

current ripples, mud-chip conglomerate, and rare current lineation and convolute lamination. Desiccation polygons are up to many tens of centimeters across. At the end of the accessible outcrop is a small interval of green sandstone, siltstone, and shale with a few HCS beds that indicate marine conditions. The upper 18 m are mostly inaccessible, dominantly shale-rich strata.

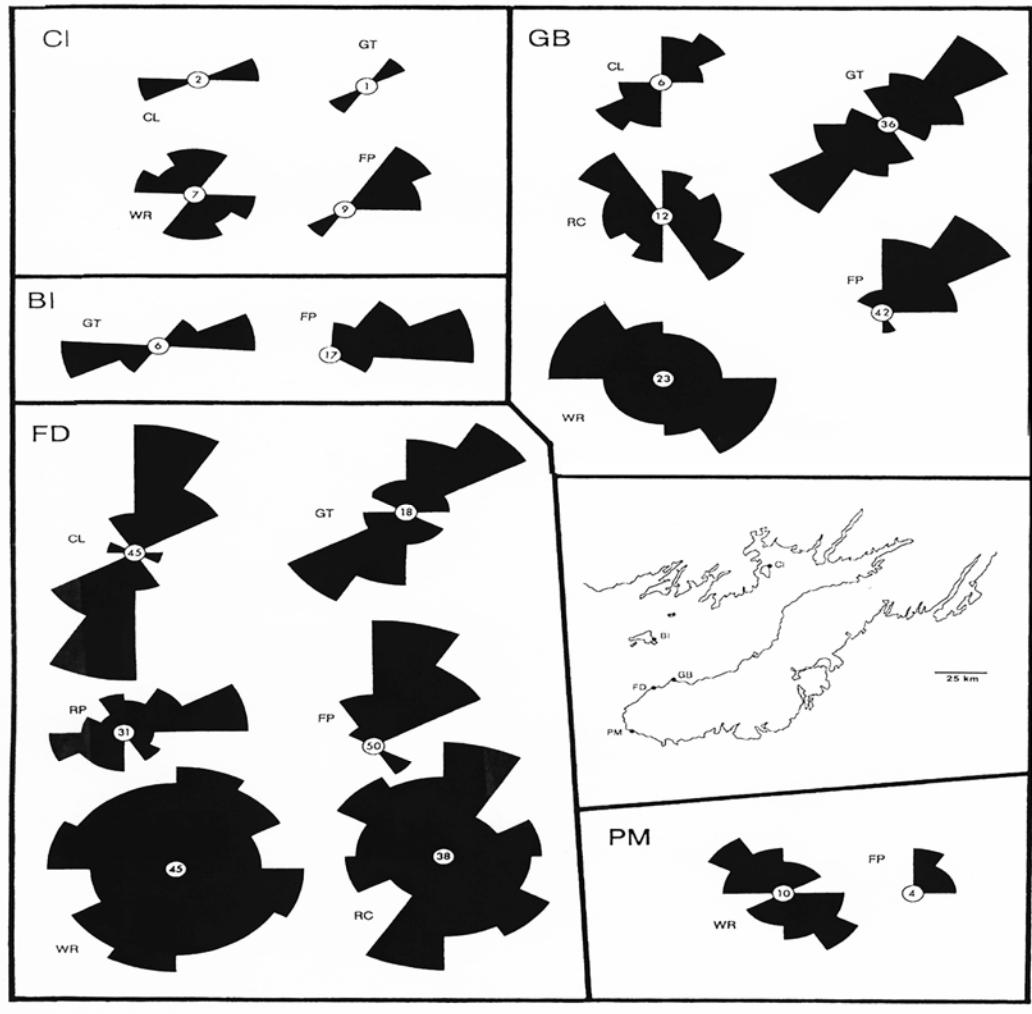


Figure 23. Paleocurrent data for member/lithofacies association 2 across Fortune Bay. CL, current lineations; GT, trend of groove marks; RP, ripple paleocurrents; WR, wave ripple crest trends; RC, ripple crest trends (type unknown). Locations: CI, Chapel Island; GB, Grand Bank Point; PM, Point May; FH, Fortune Head; BI, Brunette Island. After Myrow (1992a, fig. 16).

Interpretation

The red unit records proximal (fluvial?)–more distal (shallow marine) facies. Channelized sandstones are abundant low in the exposure and less abundant above. The vertical facies relationships indicate transition to a periodically subaerially exposed, high-energy, shoreline. The sharp stratigraphic appearance of this lithofacies suggests abrupt shallowing. No sedimentary structures are strongly diagnostic of fluvial conditions (e.g., basal lags, meter scale fining-upward cycles). No paleocurrent data or trace fossils were noted, but this does not rule out marine conditions for much or all of the facies. The shale drapes and beds and desiccation cracks in the first few meters of the fining-upward succession do not indicate deposition at the base of a distributary channel, but they could have been close to a nearshore delta plain.

Paleontology

No trace fossils were observed in this interval.

Stop 4E. Shoaling Channel Sequence at “Lighthouse Cove” (135–160 m)

Description

The rocks form low (5–10 m) sea cliffs. Redbeds form the east wall and central part of the cove. The cove’s west wall exposes a shoaling-channel succession. A thinning- and fining-upward unit (137.3–150.8 m) has lower thick–very thick sandstones and silver-green siltstone interbeds (Figure 24) that grade up into red mudstone and sandstone with many shrinkage cracks. The gray-green sandstone beds have erosional bases and tops, and show dramatic thickness changes along strike. The thicker beds are massive to faintly parallel laminated, and have amalgamation surfaces that break the beds up into divisions of 10–50 cm thick.

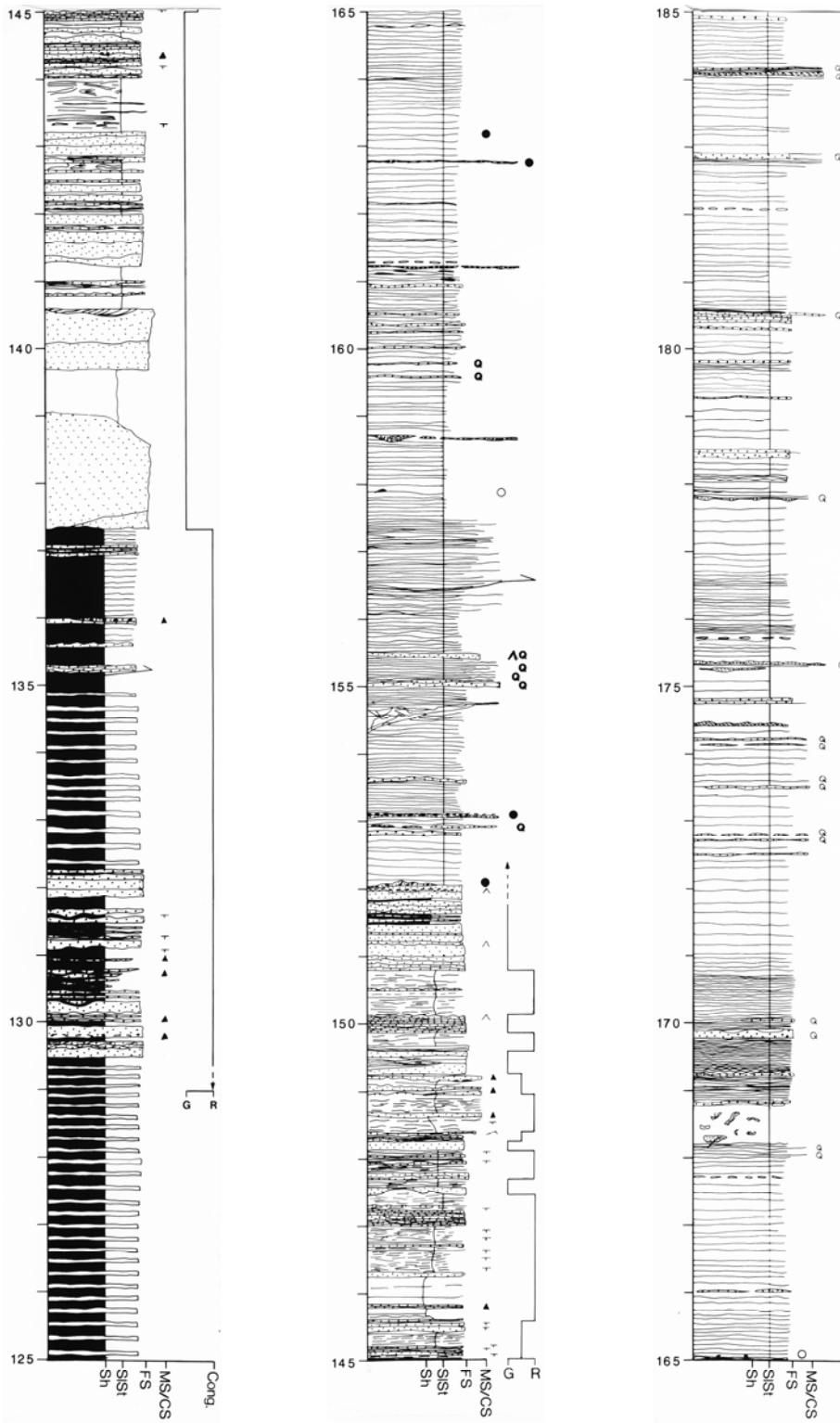


Figure 24. Detailed measured section of middle part of member/lithofacies association 2A. Section covers parts or all of Stops D–H.

Interpretation

These strata are likely subtidal deposits. They lack evidence of lateral accretion of a migrating channel (i.e., there are no thick cross-sets), but have amalgamation surfaces, indicating that deposition was in stages, but the depositional processes remain enigmatic. This shoaling-up succession possibly resulted from channel abandonment and deposition of a local mud blanket. As a rough estimate of channel depth, the thickness of strata from the base of the first and thickest sandstone bed (137.3 m) up to where the rocks take on a red tint and exhibit shrinkage cracks (145 m) is 7.7 m.

Paleontology

The thick lower package of peritidal red sandstones and shales on the south of the cove is apparently unfossiliferous. The overlying shoaling-channel sequence, both the subtidal and peritidal portions, contain trace fossils of the *Treptichnus pedum* Ichnofossil Assemblage Zone.

Stop 4F. Member/Lithofacies Association 2A (150.8–168.2 m)

Description

The lowest 1.3 m of this zone contain thick, commonly amalgamated sandstone and shale beds with abundant shrinkage cracks and mudstone clasts (Figure 24). Many upper bedding surfaces have well-aligned, spindly sphaerolite cracks. Others have large desiccation cracks, some of which record several generations of cracking. The sandy zone ends abruptly at a bed whose erosional upper surface is covered with desiccation cracks and phosphatic shale clasts. The overlying strata record a transition to the green siltstone and sandstone facies, and include several discontinuous phosphatic shale clast conglomerate beds.

Interpretation

The lowermost sandy interval may represent a period of reworking associated with maximum shoaling of the channel below. The desiccation cracks and trace fossils indicate intertidal conditions. The phosphatic shale clasts were eroded from local shoreline sources and swept into the shallow subtidal. The stratigraphic transition in this interval records deepening.

Paleontology

Psammichnites, a guide for the *Rusophycus avalonensis* Zone, first occurs approximately 130 m above the base of member/lithofacies association 2 and essentially at the same level that *R. avalonensis* Zone ichnofossils (*Rusophycus* and *Dimorphichnus*) first occur at Grand Bank Head (Stop 5E). The lower diversity at Fortune Head versus Grand Bank Head reflects fewer lower bedding surfaces in this part of the section. *Rusophycus avalonensis* occurs where soles are exposed higher in the Fortune Head section. The carbonaceous tubular fossil *Sabellidites cambriensis* occurs abundantly low in the *R. avalonensis* Zone at this locality (Figure 16C).

Stop 4G. Disturbed Strata and Raft-Bearing Bed (168.2 m)

Description

There are over 80 m of excellent lateral outcrop that show progressive deformation along the horizon (Figure 25). The horizon records a well defined change from sandstone and siltstone layers with disturbed bedding (eastern section) to a raft-bearing bed (western section). 8–10 m of cover separate these sections. The top of the bed is remarkably planar. The top of one large raft of interlaminated sandstone and siltstone, oriented at a high angle to bedding, is truncated at the bed's top. The bed shows a swirled, partly homogenized texture. In several places, laminae

directly underlie large rafts and are curved downward and compressed. Near the eastern end of this exposure, the raft-bearing bed pinches out.

Interpretation

The eastern section records largely plastic deformation, as shown by folded, rolled, and swirled laminae. The lack of a well-defined scar or slip surface (shear plane) or a consistent vergence to the folds suggest that significant downslope movement did not occur and that disruption took place by loading, possibly associated with high pore pressures and partial liquefaction. The western half of the eastern exposure (Figure 25A, B) shows that the base of the disturbed zone is better defined and may represent a shear zone. This may mark lateral transition from more-or-less *in situ* deformation to sliding or downslope movement. The relationships indicate a debris flow that was deposited by channelized flow. The rafts in the bed are clearly ripped-up pieces of previously deposited sediment. Sharp termination of the bedding in this raft clearly demonstrates that strong erosional planation formed the flat upper bed surface.

Stop 4H. Middle Member/Lithofacies Association 2A (168.2–231 m)

Description

A mass-movement deposit at 168.2 m is succeeded by 1–1.5 m of thin sandstone beds (to 20 cm thick) with eroded upper surfaces. Overlying strata are sandstones and siltstones. The lower part of this interval has many thin quartzites, and the upper has thicker sandstones.

Interpretation

This part of the succession likely records deepening from more proximal to slightly deeper subtidal settings, as recorded by the upward loss of coarser quartz sandstone beds and an

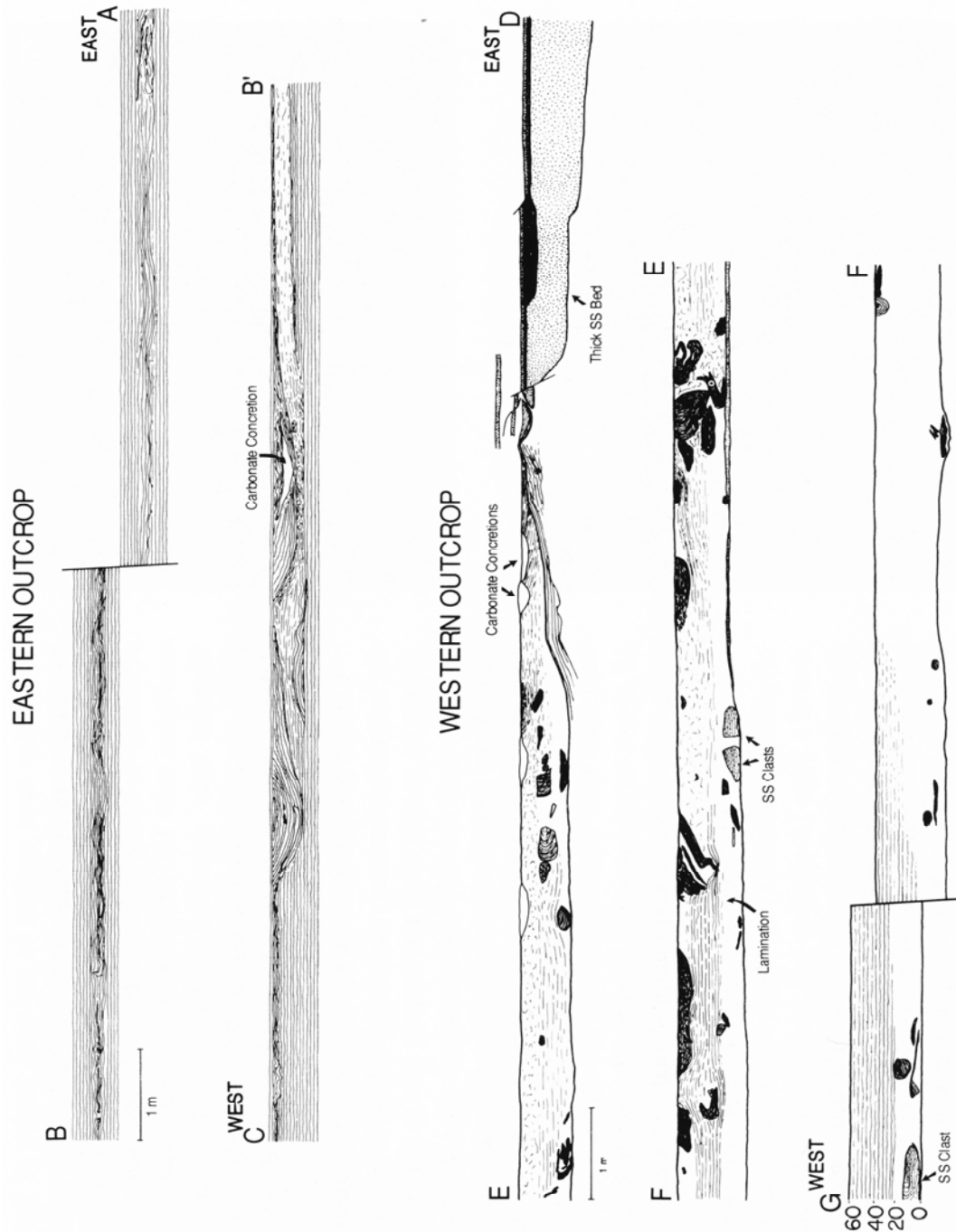


Figure 25. Detailed sketch of outcrop at 168.2 m (Stop 4G). The eastern outcrop records initial failure and disturbed bedding. Across a covered interval a few meters wide, the western outcrop shows details of a debris flow deposit (debrite) and sandstone channel. From Myrow and Hiscott (1991, fig. 7).

upward increase in sandstone bed thickness. The sandy zone at the base of the interval likely records temporary shoaling. The features at the top of the raft-bearing bed at 223.6 m indicate postdepositional loading of overlying strata (and possible later exaggeration by compaction).

Paleontology

Simple grazing trails (*Helminthopsis tenuis*, *Helminthoidichnites tenuis*, *Cochlichnus anguineus*) and burrows such as *Planolites* are abundant. *Treptichnus pedum*, *Psammichnites*, and *Monomorphichnus* occur sporadically through this interval. Microbially covered surfaces are well exposed in the interval 221–225 m and are associated with *Monomorphichnus* and a variety of grazing trails preserved on upper bedding surfaces (Buatois *et al.*, 2014).

Stop 4I. Slumped Horizon and Overlying Strata (231–245 m)

Description

At 231.1 m, a 1.7 m-thick horizon consists of a lower chaotic unit 30–50 cm thick, and an upper siltstone unit 1.2–1.4 m thick (Figure 26). The upper siltstone unit lacks macroscopically recognizable laminae. The chaotic lower unit consists of a 0–19 cm-thick sandstone overlain by large, variably oriented sandstone slabs within deformed interbeds of sandstone and siltstone. The lower sandstone bed shows rapid changes in thickness, partly as a result of basal erosion. Parts of it appear massive, while others display contorted laminae, ball-and-pillow structures, and irregularly shaped clasts and oblate spheroids of carbonate-cemented sandstone. In the 10–15 m above the slump horizon, there are many thicker sandstone beds (8–15 cm) with erosional tops and bottoms and a few hummocky cross-stratified beds.

Interpretation

The horizon at 231.1 m records sliding on a truncation surface. Distinctive laminae in the sandstone blocks indicate they were derived from the associated sandstone bed and then rotated within the mass as it moved downslope. The lower sandstone bed shows evidence of flow. The overlying siltstone is an unusually thick, massive bed, which, along with its lack of internal structure or grading, suggests a mudflow origin. The two parts of the bed are genetically related. The interval records failure of the sea floor, including initial sliding and a subsequent mud flow. The thicker, HCS sandstone beds above are consistent with slightly more distal deposition than the thinner bedded sandstone and siltstone facies.

Stop 4J. Upper Part of Member/Lithofacies Association 2A (245–285 m)

Description

This interval has interbedded sandstone and siltstone. There are many medium beds of sandstone, 10–27 cm-thick, which have highly erosional tops. Wave rippled upper surfaces are common. A few thin quartzitic pebble/granule horizons (e.g., 250.2 m) and phosphatic shale clast beds were noted. There are several raft-bearing beds at this stop as well.

Interpretation

This interval is sandier than most parts of member/lithofacies association 2, and has more evidence for post-depositional reworking of tempestites. Mass-movement deposits are less abundant, and quartzitic beds are virtually absent. We interpret this interval to be a more distal inner shelf deposit where there was less throughput and more deposition. The lack of

disorganized beds reflects the more distal depositional setting where the sedimentation rate of mud was reduced relative to the very nearshore environments.

Paleontology

Trace fossils are abundant in this interval and, in addition to *Planolites*, include *Treptichnus pedum*, *T. coronatum*, *Gyrolithes*, *Psammichnites*, *Helminthopsis abeli*, and *Cochlichnus*.

Monomorphichnus is common throughout the interval. At least 16 matground surfaces

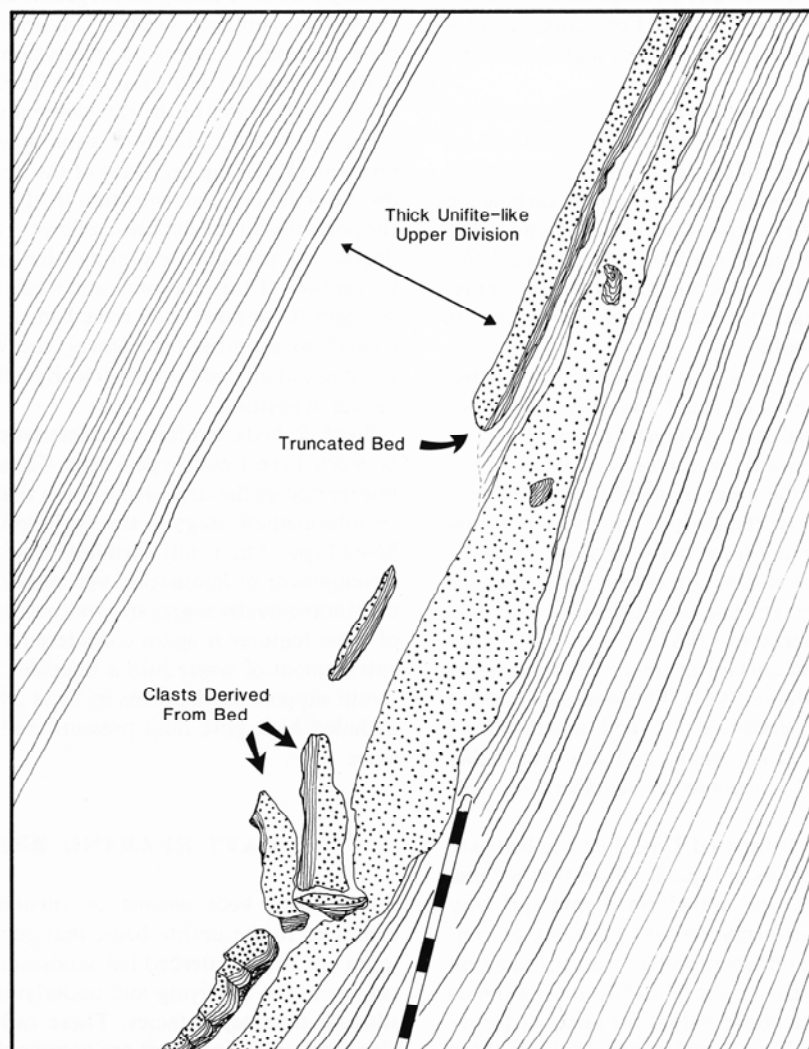


Figure 26. Sketch of thick mud flow deposit with truncated bedding and blocks at base at Stop 4I. From Myrow and Hiscott (1991, fig. 11).

occur in this interval (Buatois *et al.*, 2014). Significantly, the arthropod burrow *Rusophycus avalonensis* (Figure 19A) and the arthropod walking trail *Allochotichnus* are preserved in epirelief on matground surfaces (Figure 19C–E).

Stop 4K. Base of Member/Lithofacies Association 2B (285–312 m)

Description

A 5.2 m-thick unit of red siltstone with thick laminae along with very thin beds of white to green, fine to medium sandstone (ca. 30% volume of rock) marks the base of member/lithofacies association 2B. The red siltstone contains a small percentage of fine- to medium-grained quartz sand and detrital micas (muscovite and minor biotite). The bedding ranges from regular to extremely discontinuous. There are no shrinkage cracks, but there are small starved current ripples and thin beds with wave-generated lamination. The top of this interval (including a 90 cm-thick red horizon) is exposed farther to the west, just beyond a fault zone (Figure 11).

This red bed is overlain by approximately 25 m of green siltstone with very thin and thin, very micaceous fine sandstone beds (<5% of rock) that show few bedding plane or sole features and a paucity of trace fossils. The sandstone beds make up less than 5% of the rock; the most sandstone-deficient exposures contain as little as 1–2% sandstone and lack sandstone beds >1.5 cm thick for as much as 5 m through the section. The sandstone beds display parallel lamination and ripple cross-lamination, with the latter in some cases stacked vertically. Pyritic, phosphatic, and carbonate nodules exist as single, isolated features or in trains along thin sandstone beds. The phosphatic nodules are dominantly elongated parallel to bedding, and average approximately 2–3 cm in length and 1 cm in thickness.

Interpretation

Both the red and green siltstone units were deposited in quiet-water environments with low sand input. Their extensive areal distributions, along with their gradational relationship with overlying and underlying deltaic deposits, suggest that these strata are an extensive fine-grained blanket formed during delta abandonment (Myrow and Hiscott, 1993). For the red siltstone, the fine grain size, nature of the bedding, and small scale of sedimentary structures indicate that the sediment was deposited under generally low-energy conditions in which gentle currents and waves reworked the bottom. The red color, lack of shrinkage cracks, and presence of burrows imply oxygenated, but subaqueous, deposition.

The sediment of the green siltstone unit was not disturbed by infauna, which along with the presence of pyrite nodules, suggests that the interstitial fluids, and possibly even the bottom waters, may have been poorly oxidized during deposition. Episodic deposition of sand by currents and waves led to thin stratified beds. During abandonment, continued subsidence and reduced sedimentation may have also caused deepening.

Paleontology

Large circular burrows 1–3 decimeters in diameter occur at several levels low in member/lithofacies association 2B.

Stop 4L. Member/Lithofacies Association 2B (312–442m)

Description

The interval contains sandstone and siltstone facies that at the upper limit of the outcrop transition into member/lithofacies association 3, which is exposed in faulted outcrop farther

down the shore to the west. There is a well developed HCS bed at 333 m and a concentration of HCS beds between 365 and 375 m. These fine- and very fine-grained sandstone beds show meter-scale hummock spacings, amalgamation surfaces, and in some cases, a lower division of structureless and/or parallel lamination. Symmetrical ripples mark the bed tops.

A chaotic/deformed horizon is exposed at 346.2 m. It contains 35–40 cm of thin and very thin beds of sandstone and siltstone that terminate against a westwardly downcutting shear zone/surface that is overlain first by folded strata and then by cross-stratified fine sandstone. To the west, the surface is bedding-parallel and forms a well-defined decollement. The overlying strata are regularly folded into sharp anticlines and wide, gentle synclines that are filled with large cross-laminae of fine sandstone. The overlying sandstone bed contains extensive, low-angle laminae that form widely spaced, low relief, convex-up domal patterns typical of HCS. The upper surface of the sandstone bed has symmetrical-crested ripples. Overlying this deformed interval is 3–4 m of strata with anomalously high sandstone content (ca. 90%). This includes beds, many of which are amalgamated, that are up to 25 cm thick. Erosional upper surfaces are common, as well as parallel lamination, HCS and wave ripples.

This stop contains only one disorganized bed—a 50 cm-thick raft-bearing bed at 360.15 m with a remarkably flat base and large rafts. The lower part of the bed appears very homogeneous and the upper part may have some cryptic internal lamination.

Interpretation

The majority of this stop shows thin- and medium-bedded tempestites, including HCS beds, and a paucity of disorganized beds. These are interpreted as deeper shelf deposits formed in the zone in the tempestite depositional model in which the throughput-to-deposition ratio is at

a minimum (Figure 22). This area of the shelf was characterized by large swells (deposition of HCS) and lower mud depositional rates (fewer mass-movement deposits). The upper parts of the stop record a transition to the more distal shelf deposits of member/lithofacies association 3.

The buckled horizon at 346.2 m records compressional stresses (oriented northwest), and both ductile and brittle deformation. The features of this horizon suggest a close association of sand deposition with sliding. The overlying HCS sandstone implicates storm processes in the sediment failure. Storm-generated waves are a factor in many slope failures due to changes in pressure associated with the passing of waves. The resulting cyclic shear stresses decrease sediment strength and increase pore pressure, which leads to failure and downslope movement.

Paleontology

Trace fossils are common in the gray-green sandstone and siltstone facies. Several matgrounds are present, one of which exhibits the arthropod scratch marks *Monomorphichnus* and the arthropod walking trace *Diplichnites* on its upper surface. Several large bedding surfaces spectacularly covered with *Psammichnites* isp. occur at the western end of the outcrop.

These *Psammichnites* specimens show guided meandering patterns, including strophotactic, phobotactic, and thigmotactic behaviors that signal the appearance of the Cambrian information revolution (Mángano and Buatois, 2014, 2016; Carbone and Narbonne, 2014). The lowest known shelly fossils in the Chapel Island succession, decalcified and sediment-filled conchs of the orthothecid "*Ladataeca*" *cylindrica*, are found in green siltstones interbedded with thin green-grey sandstones approximately 415.9–419.5 m above the base of member/lithofacies association 2 (Landing *et al.*, 1989), which is at ~400 m on the section log. A

diagrammatic representation of the biological changes from the Ediacaran to the Fortunian is shown in Figure 27 (from Buatois *et al.*, 2014).

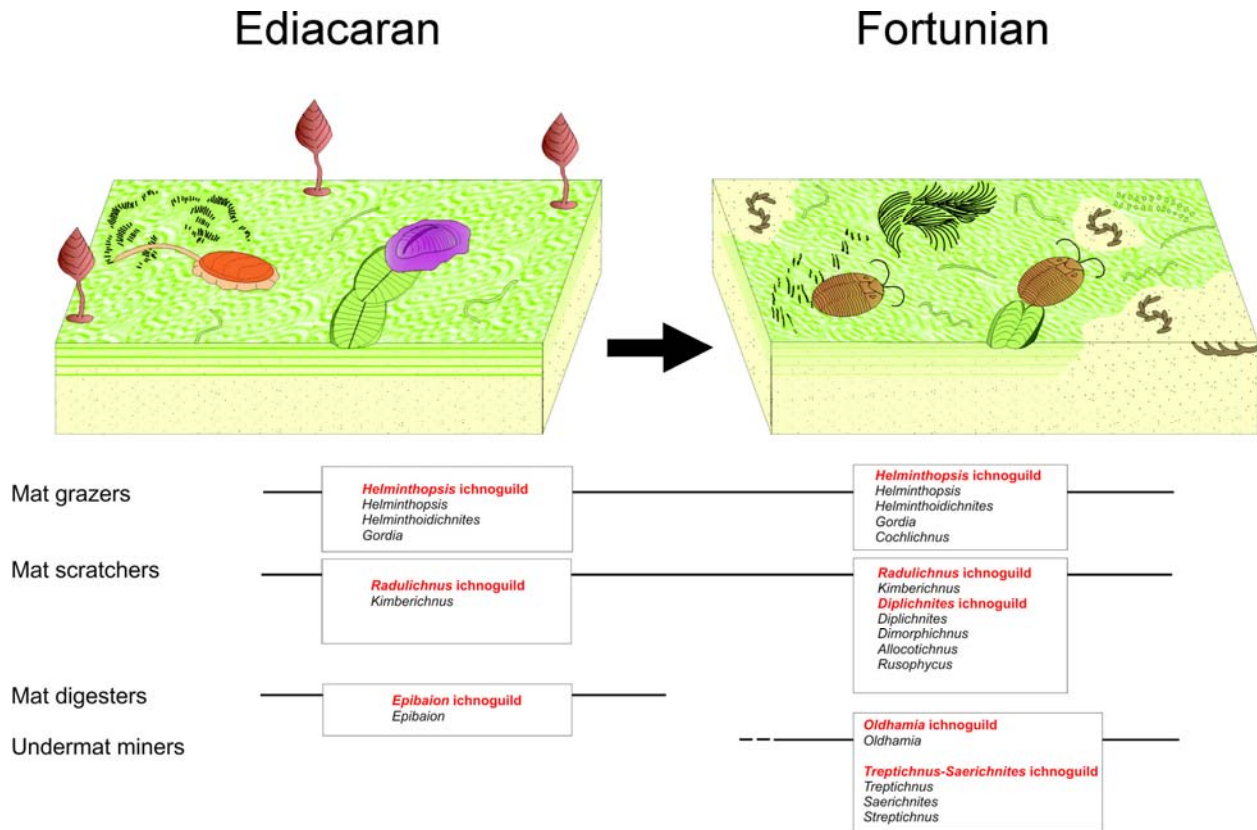


Figure 27. The ichnological record of animal–matground interactions across the Ediacaran–Fortunian transition (after Buatois *et al.*, 2014).

Stop 5. Grand Bank Head: Ediacaran–Cambrian Boundary Reference Section (June 25, PM, GMN, MGM, LAB, RG, BL, EL)

Location

The locality is on the west side of Grand Bank village. Drive down College Street to the NW side of Admiral Cove. Park near the end of the street at the sharp left. A section from the upper Rencontre Formation through the lower Chapel Island Formation (members/lithofacies

associations 1–2B) is exposed around Grand Bank Head (Figures 28, 29). Ca. 400 m of strata will be examined (Figure 30), with the outcrop divided into three parts separated by covered intervals (GBA, GBB, and GBC). The outcrop consists of steep sea cliffs separated by large coves. The lower and upper parts of the section are easily accessible, while the central part (Stops 5D and 5E) is more difficult to reach (Figure 29). The strata are vertical or steeply dip ($>80^\circ$) northwest. The section is disrupted by numerous faults, most of which have small offset.

Stop 5A. Upper Rencontre Formation

Description

Over 100 m of strata are exposed along the shore on the north side of the cove. Red, micaceous, fine grained, thinly laminated sandstone, siltstone, and mudstone exhibit a wide variety of sedimentary features that include ripple marks, shrinkage cracks, intraclast conglomerate, and ball-and-pillow structures. Thin- to medium-bedded sandstone beds contain parallel lamination, ripple cross-lamination, scour-and-fill features, and clay-draped scour surfaces.

Interpretation

Depositional environments for the lower Rencontre Formation were likely alluvial/fluvial, but for the middle and upper parts, particularly near this transition into the Chapel Island Formation, they were more likely marginal marine or peritidal.

Fossils

Despite vigorous search, no fossils have been found.

Stop 5B. Lowest Member/Lithofacies Association 1 of the Chapel Island Formation

(GBA-0–50 m)

Description

Red and green sandstone, siltstone, and mudstone make up the base of the formation. The facies consists of very thin–medium beds of sandstone (roughly 50–80%) with thinner interbeds of siltstone and shale. Flaser, wavy, and lenticular bedding are common. The sandstones have sharp lower surfaces and flat, rippled, or very irregular upper surfaces; are structureless, parallel-laminated, or ripple cross-laminated; and are commonly amalgamated. A common motif in these sandstone beds consists of the following: (1) a lower division of structureless sandstone with abundant angular shale clasts, overlain by (2) parallel-laminated sandstone with or without scattered shale clasts, capped by (3) current ripple cross-lamination. Erosional surfaces are abundant in this lithofacies, in places, and define isolated, lenticular, scour-and-fill or channel-fill beds of sandstone or interlaminated shale and sandstone (Figure 31).

Sedimentary structures on upper bedding surfaces include straight, sinuous, and interference ripples and rare “raindrop” prints. Sandstone-filled desiccation cracks (polygonal patterns) and syneresis cracks (plan view: three-armed star shapes or linearly-oriented, straight to slightly irregular spindles) are common. Paleocurrent data (n=58) from current ripples (Figure 32) indicate bimodal-bipolar flow conditions, with the stronger mode towards the southwest.

Interpretation

Many of the features of this facies are found in fluvial environments (e.g., rip-up clasts, desiccation cracks), but the full suite of sedimentary structures and overall organization of beds — there are no meter-scale, sandy, upward-fining succession with facies typical of meandering

rivers — indicates marine conditions, as does the, albeit, limited suite of trace fossils. Many of the structures are well known from modern and ancient shallow-water, tidally influenced facies, although these structures are certainly not restricted to tidal settings.

Thinly interlayered bedding, ubiquitous mud-chip conglomerate, abundant scour surfaces, desiccation and syneresis cracks, bimodal–bipolar paleocurrent distribution, and other features are most consistent with a tidally influenced peritidal environment. The bedding style and associated structures would be most typical of lower tidal flat to shallow-subtidal settings.

Paleontology

Despite vigorous search, no definite fossils have been found.

Stop 5C. Member 1 of the Chapel Island Formation (GBA-65–120 m)

Description

The slope leading down into the cove is steep, and access may be difficult if the rocks are wet. Steeply dipping strata from 65–97 m are exposed on the south side of the cove. A fault/shear zone separates the lower part of the section from the strata on the north side of the cove (97–120 m).

65–97 m Zone: The strata are similar to those of the last stop, but also contain shale-rich (50–80%) intervals. The latter show wavy and lenticular very thin beds and laminae of sandstone; thin and medium beds of sandstone make up only a small percentage (<10%) of this lithofacies. Parallel-laminated beds, shale-clast conglomerate, erosional surfaces, and channel sandstone beds are less common. Starved ripples and shrinkage cracks are particularly abundant.

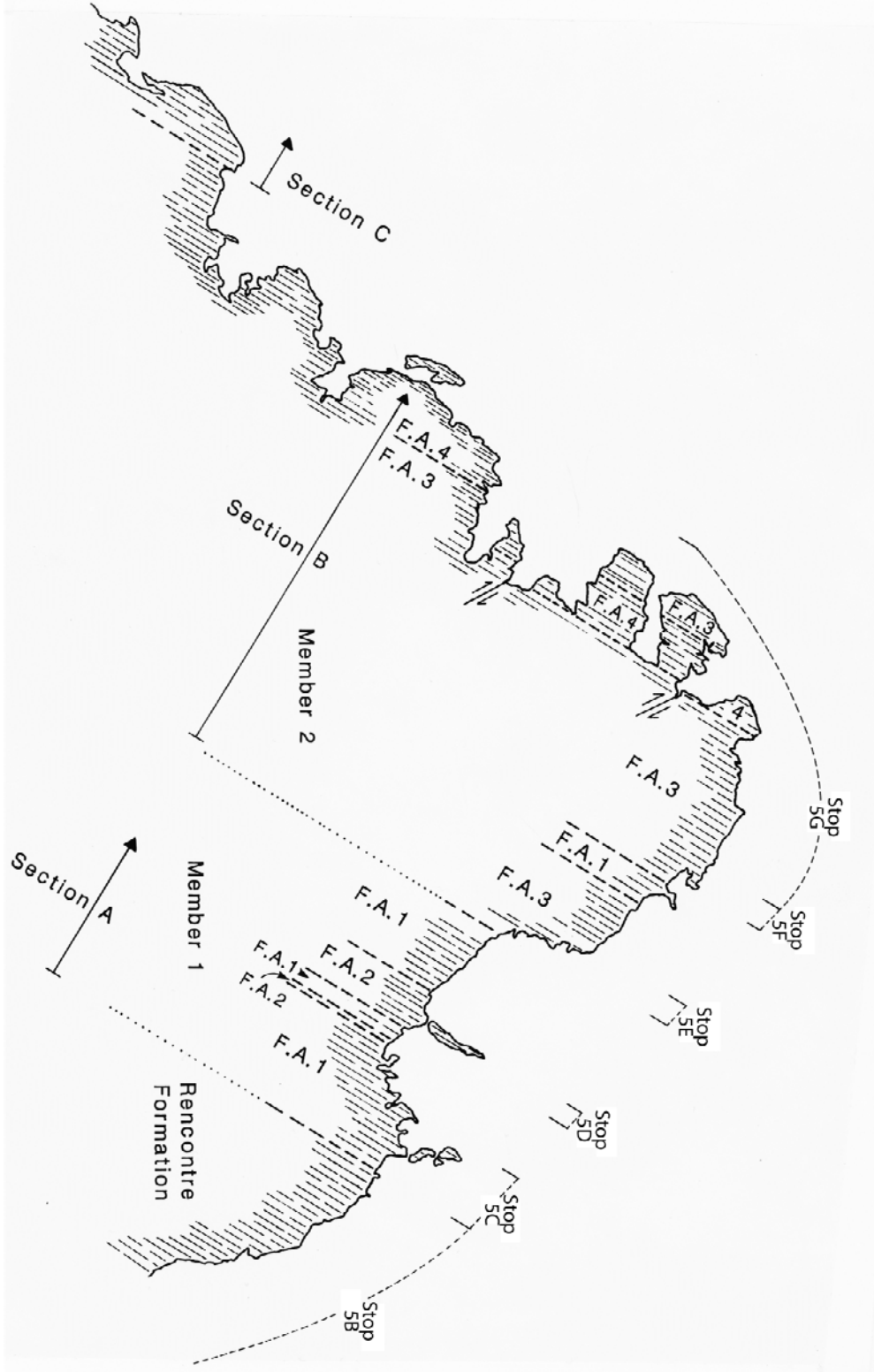


Figure 28. Grand Bank Head, dashed lines indicate strike of strata. Magnetic north to right of figure; scale 3 cm = 100 m.

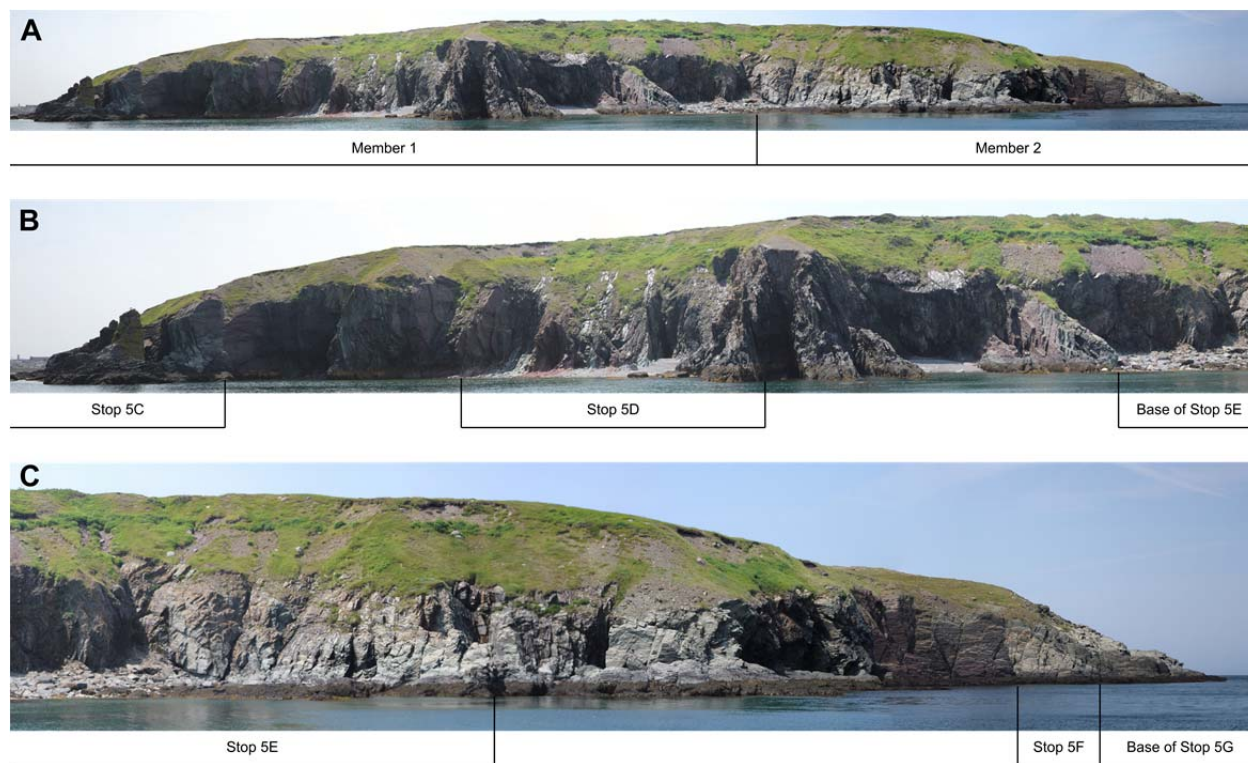


Figure 29. *Panorama of the section at Grand Bank Head (Stop 5). Contact between member/lithofacies association 1 and 2 visible in all three images. Panorama images provided by Huan Cui. A, View of the entire headland shows the distribution of member/lithofacies association 1 and 2 of the Chapel Island Formation. B, View of member/lithofacies association 1 of the Chapel Island Formation shows locations of stops 5C, 5D, and base of 5E. Intervening strata are only readily accessible by boat. C, View of member/lithofacies association 2 of the Chapel Island Formation shows locations of stops 5E, 5F, and base of 5G. Intervening strata only readily accessible by boat.*

The association of shrinkage cracks with the color of the strata is very pronounced: red units normally have very abundant shrinkage cracks, while green units are more variable with some having a moderate number of shrinkage cracks, and others having few, if any. A thin package of sandstone-laminated black shale is also present in this zone (77.9–78.2 m).

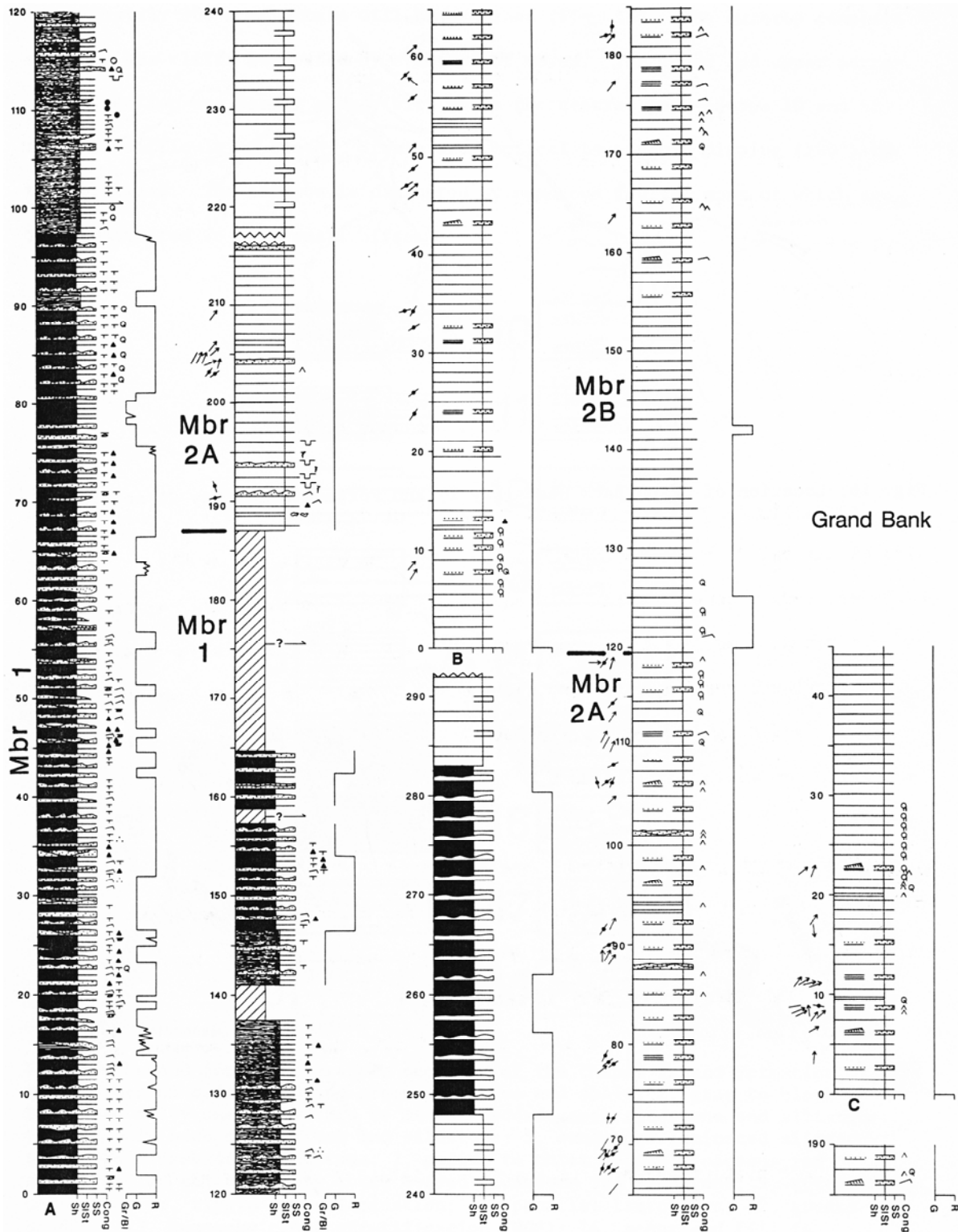


Figure 30. Stratigraphic section for Grand Bank Head. Symbols to left of column are paleocurrents with north at top of page. Color is given by vertical lines to right of column (Gr/Bl, gray/black; G, green; R, red).

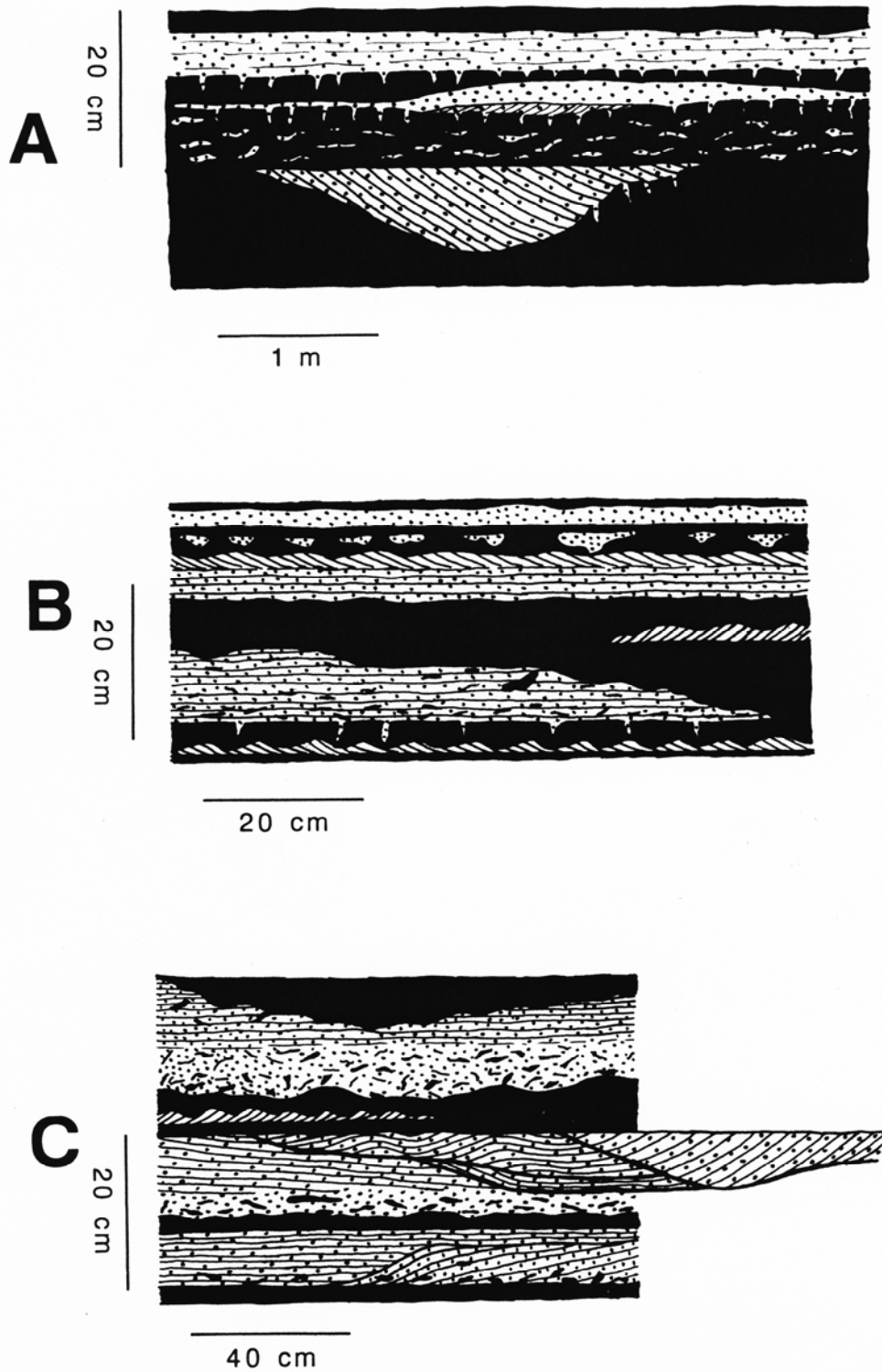


Figure 31. *Sketches of bedding in member/lithofacies association 1 of the Chapel Island Formation at Grand Bank Head. Note channel bodies and shrinkage cracks.*

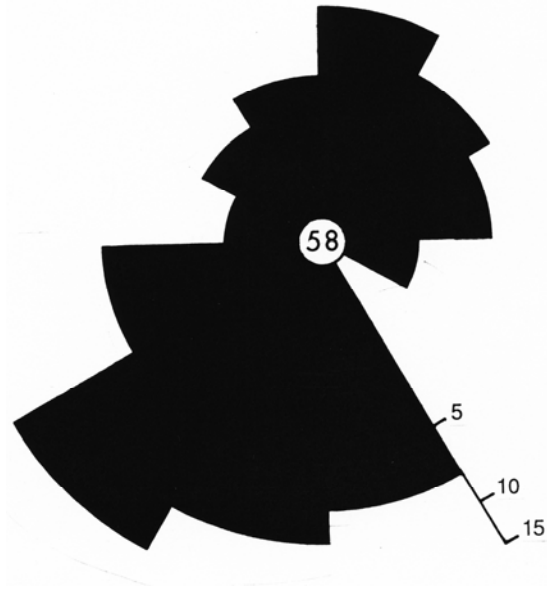


Figure 32. *Equal-area rose diagram for current ripples ($n = 58$) at Grand Bank Head.*

97–120 m Zone: The strata exhibit a complex interbedding of laminated dark shale and silver-grey siltstone with thin to medium-bedded white quartz arenite and gray-green sandstone beds. There is also less abundant, very thinly laminated (pin-striped), dark gray to black shale with a 40–60% content of very fine- and fine-grained, white, quartzose sandstone laminae. The silver-gray siltstone is interbedded with the shale on a lamina-by-lamina scale or on a larger scale in which centimeter-scale siltstone beds alternate with centimeter- to decimeter-thick beds of pin-striped shale. Beds of quartz arenite and sandstone range from thin starved ripples to medium beds with channel-like geometries to those with relatively flat bases and eroded tops. This lithofacies contains abundant soft-sediment deformation structures, including small-scale slides, convolutions, and ball-and-pillow features. Some samples contain small (1–3 mm long) injection structures (clastic dikes) and small-scale syndimentary faults. Synaeresis cracks are locally abundant, as are early-diagenetic pyrite and phosphate nodules.

Interpretation of 65–97 m Zone

The higher shale/sandstone ratio and lower average sandstone bed thickness indicate lower-energy conditions. The paucity of large-relief erosional surfaces and the reduction in abundance and scale of small scours and channels mean that current energy was mostly below that sufficient to erode consolidated mud. Given lower-energy currents, less channelling, and more extreme and frequent desiccation, this interval was possibly deposited in shallower water than the last stop, perhaps in a middle to upper tidal flat setting.

Interpretation of 97–120 m Zone

The laminated shale and siltstone facies at this stop stratigraphically alternate with the peritidal facies described above, which, along with the abundance of synaeresis cracks, support a shallow-water interpretation. The shale, with its lenticular/wavy bedding and incipient ripple cross-laminae, indicates low energy suspension deposition of sand and mud with gentle, episodic current activity. The sandstone beds record higher-energy deposition. The abundance of soft-sediment deformational structures indicates high pore pressure conditions, probably linked to high sedimentation rates and the presence of liquefaction-susceptible silt. High sedimentation rates would have also favored high organic content, by shortening the time in which organic matter remained in the upper, oxidizing diagenetic zones. The organic matter would have favored reducing conditions and the precipitation of phosphorus and pyrite nodules. Whereas low oxygen conditions are indicated, dysaerobic rather than anaerobic conditions are considered more likely because of the presence of a moderate abundance of trace fossils.

These facies are consistent with restricted, low-oxygen, nearshore environments such as estuaries, tidal-flat ponds, and interdistributary bays of deltas. Interdistributary bays can be very

extensive and contain thick deposits, while maintaining generally shallow-water, restricted conditions. In this interpretation, the medium quartz arenite and sandstone beds could be both channel-fill and thin crevasse-splay deposits.

Paleontology

The vendotaenid alga *Tyrasotaenia* is locally abundant in black shales and siltstones near the top of this interval. The horizontal burrow *Helminthoidichnites* is relatively common and the serial Ediacaran megafossil *Harlaniella podolica* occurs rarely in these beds.

Stop 5E. Lowermost member/lithofacies association 2A (GBA-187–215 m)

Description

The strata are stratigraphically above a faulted and partly covered interval that includes the member/lithofacies association 1–2A boundary. The rocks consist of silver-green siltstone with very thin to thin beds of gray-green sandstone that form a gutter cast facies. Sandstone dikes and pot and gutter casts are abundant. The bedding style and sedimentary features are the same as that described for Stop 4B.

Interpretation

Based primarily on the lithofacies relationships and the types of sedimentary structures, Myrow et al. (1992ab) interpreted this lithofacies as the deposits of a shallow marine setting along a storm-influenced shoreline. Storm-generated flows cut the pot and gutter casts and deposited the thin sandstone beds in a nearshore zone of sediment bypass.

Paleontology

Well-preserved specimens of the trace fossil *Treptichnus pedum* along occur in thin sandstones near the base of member/lithofacies association 2 along with specimens of *Planolites*, *Cochlichnus*, anemone burrows, and specimens attributed to *Monomorphichnus*. In 2013, a single specimen of *T. pedum* was found immediately below the base of member 2 in a loose slab that most closely resembles that of subtidal beds in the uppermost meter of member/lithofacies association 1 or the basal 2 m of member/lithofacies association 2.

In general, the succession of strata and trace fossils in uppermost member/lithofacies association 1–basal 2 at Grand Bank Head is remarkably similar to the succession that we saw in the basal beds of the GSSP at Fortune Head (Stop 4B).

Stop 5F. Part of member/lithofacies association 2A (GBB-0–35 m)

Description

The strata consist of silver-green siltstone with very thin to thin beds of gray-green sandstone. This facies contains anywhere from 5–40% sandstone in very thin to thin beds. The rocks of this facies at this locality show unusually pervasive burrowing.

Interpretation

As described for Fortune Head, the sandstone beds are thought to represent thin tempestites deposited in shallow subtidal environments. Gravity flow structures are not present in this interval, but occur in the overlying strata and at the equivalent stratigraphic level in the Fortune Head section. The facies was deposited under conditions of high sedimentation rates probably within a deltaic setting.

Fossils

Trace fossils are unbelievably abundant on most bedding soles. Specimens of *Dimorphichnus* and very shallow (less than 2 mm-deep) *Rusophycus avalonensis* (Figure 17G) appear for the first time in the Chapel Island succession and define the base of the *Rusophycus avalonensis* Zone. As discussed above in this field guide, *Rusophycus* from member/lithofacies association 2 of the Chapel Island Formation provides the oldest unequivocal evidence for the existence of crown-group Ecdysozoa, Lobopodia, and Arthropoda in Earth's biotic evolution (Benton *et al.*, 2015; Wolfe *et al.*, 2016).

The bedding soles are nearly totally covered by trace fossils, mostly branching burrows showing affinities with *Phycodes*, possible preservational variants of *Psammichnites*, *Rusophycus avalonensis*, and other trace fossil taxa (Crimes and Anderson, 1985; Narbonne *et al.*, 1987; Mángano and Buatois, 2016). These surfaces show a high bedding-plane bioturbation index (BP-BI) for the first time in the Fortunian (Mángano and Buatois, 2016). Despite the high density of trace fossils, they are essentially parallel to bedding planes and create little vertical disturbance of the primary fabric, a style of animal activity that is typical of the Fortunian (Mángano and Buatois, 2014).

Stop 5G. Bulk of member/lithofacies association 2A (GGB-35–120 m)

Description

This stop covers a long stretch of coast that extends towards the end of the headland. This interval ends at a thick conspicuous red unit described at the next stop. The lower and upper parts of this interval consist of silver-green siltstone and very thin–medium, gray-green sandstone beds. These intervals have a higher percentage of sandstone (40–60%), and many beds that are

greater than 3 cm in thickness (~10–15%). The thicker sandstones show grading and well-developed sole marks that indicate consistent northeast-directed paleocurrents. Many are parallel laminated at their base and contain oscillatory- and combined-flow lamination higher in the bed.

There are many disorganized beds at this stop that include a full range of types from mudflow beds to raft-bearing debrites. An anomalously thick (1.1 m) mudflow bed is present at 74.3 m, and an 80 cm-thick raft-bearing bed is at 113.8 m. The latter has a 15–20 cm-thick, hummocky cross-stratified sandstone bed directly above it. The association of a mass-movement deposit with a storm-generated feature is also found at 346.2 m at Fortune Head, and implicates storm-generated cyclic wave loading as a contributor to slope failure.

Interpretation

Deposition in nearshore and inner-shelf settings is based on close association with shoreline facies (with evidence of episodic exposure), HCS, and abundant wave ripples and wave-ripple laminae. The subtidal and inner shelf deposits contain considerable muddy siltstone, which on the modern inner shelf is generally associated with large deltas, especially with moderate–low wave and tidal influence. These deltas often have high fine-grained sediment deposition rates (e.g., Mississippi and Huanghe (Yellow River) deltas). Abundant instability features (disorganized beds) require high sedimentation rate to yield conditions that allow failure (high pore pressures and liquefaction) on gentle nearshore marine slopes.

Paleontology

Trace fossils occur commonly throughout the interval. They include *Treptichnus pedum*, *Monomorphichnus* (including specimens preserved on upper surfaces in matgrounds), and small

Gyrolithes, *Treptichnus pedum*, *T. coronatum*, *Helminthopsis abeli*, *Helminthoidichnites tenuis*, *Diplichnites*, and *Rusophycus* are especially common at the top of this interval immediately below a distinctive interval of red and green siltstone (Facies 4). All of these strata are part of the *Rusophycus avalonensis* Zone.

Stop 5H. Lower Part of member/lithofacies association 2B (GBB-120–150 m)

Description

This stop exposes the same red and green laminated to very thinly bedded siltstone units described at Stop 4K. The 5.2 m of red siltstone horizon at 120 m defines the base of member/lithofacies association 2B.

Interpretation

See detailed interpretation of Stop 4K.

Fossils

The trace fossils are similar to those described at Stop 4K.

Stop 5I. Member 2B (GBB-150–190 m)

Description

Continue on from Stop 5H to the end of the headland (approximately 165 m). From this point onward along the coast to the southwest, the outcrop is offset by numerous faults. One can drop into coves and pick up marker horizons and measure approximately to the 190 m mark before losing continuity of section. The strata at this stop show a return to sandstone and siltstone facies, which are described at Stops 4C and 4L.

Interpretation

See Stop 4C for interpretation of the facies.

Fossils

The trace fossils are similar to those of Stop 4L, and include *Treptichnus coronatum*, *Helminthopsis abeli*, and *Psammichnites* isp., the latter of which shows complex meandering behaviour.

Stop 6A. Little Dantzic Cove: Stage 2/“Laolinian Stage” (June 26; EL, PM, GM, GMN, LAB, RG, BL)

Location

Little Dantzic Cove (Stop 6A; Figure 2, LDC) and Little Dantzic Cove Brook (Stop 6B) are adjacent coastal sections. Drive ca. 23 km S of Fortune village on Route 220. Turn E on the (very unimproved!) dirt road to Pieduck Point; park near the wharf and walk north along the cliff top. The section at Little Dantzic Cove Brook is 3.8 km N of the parking lot, and the Little Dantzic Cove section is 4.5 km north of the wharf. The section of the Chapel Island Formation at Stop 6A includes over 400 m of strata (Figure 33). Two sections in the Random–Chamberlain’s Brook formation interval are exposed in synclines just north of Pieduck Point, as well as 0.5 km to the north at Snooks Brook, and further N at the Little Dantzic Cove Brook section.

Generalized Stratigraphy and Significance

Stops 4 and 5 record the earlier part of the Cambrian Evolutionary Radiation (CER) by recording the ichnofossil succession and its significance on substrates from the terminal

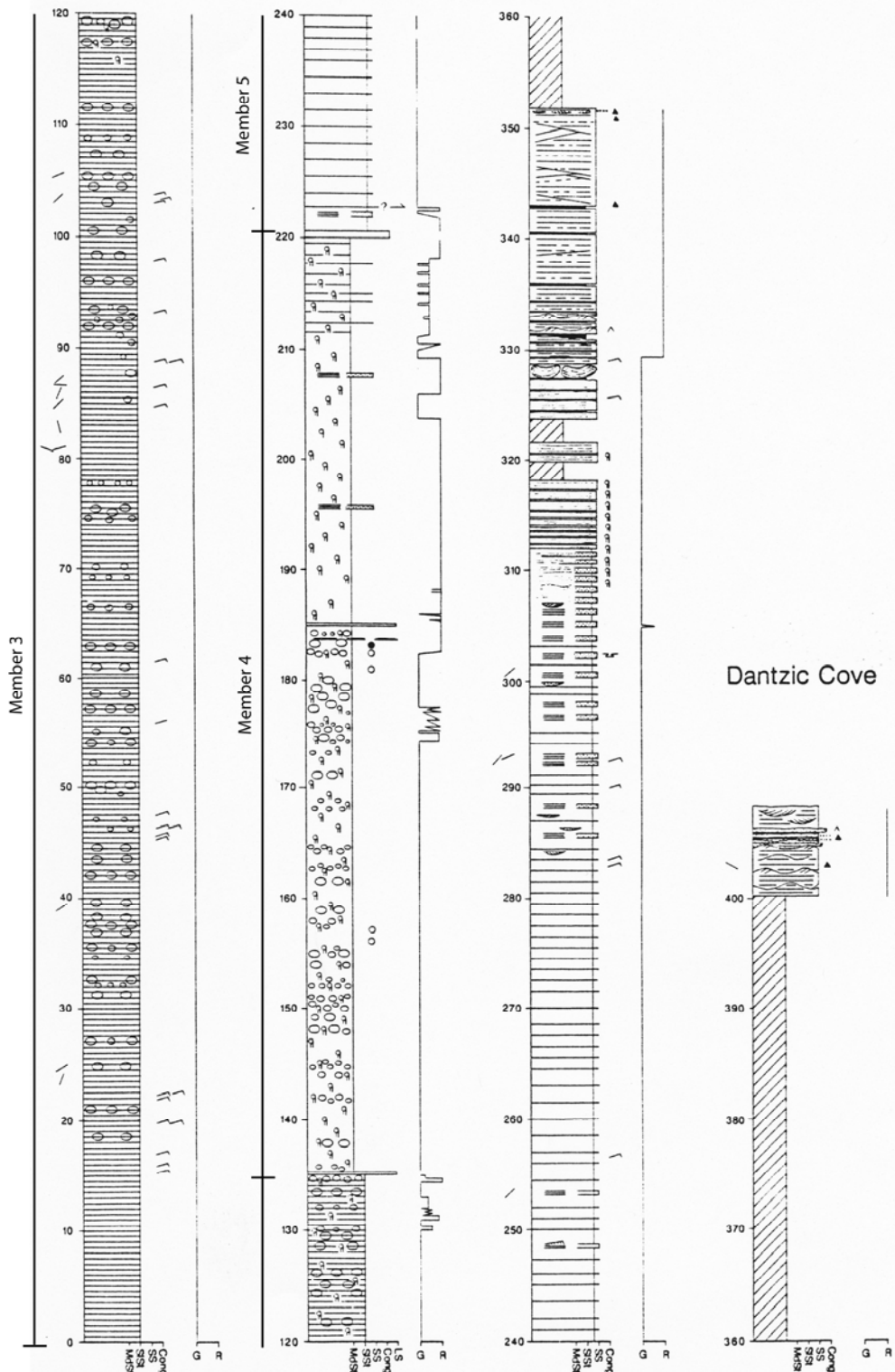


Figure 33. Stratigraphic section for Little Dantzig Cove. Symbols to left of column are paleocurrents with north at top of page. Color is given by vertical lines to right of column (Gr/Bl, gray/black; G, green; R, red).

Ediacaran through most of the lowest Cambrian Fortunian Stage in the Rencontre–middle Chapel Island formations of Avalonia. A monospecific, biomineralized fauna (the orthothecid “*Ladatheca*” *cylindrica*) appears in the middle Chapel Island Formation (Stop 4L) as a harbinger of the diversification of a biomineralized biota. This second stage of the CER and beginning of Stage 2/the proposed Laolinian Stage is distinguished by the shallow-water appearance of the stratigraphically long-ranging *Watsonella crosbyi* Zone fauna (ca. 10 m.y.) that included benthic conoidal and molluscan taxa. In addition, the oldest mid-water predators with weakly to strongly mineralized grasping spines (*Protohertzina*) appear in the upper middle Chapel Island Formation a relatively short distance below a carbon isotope excursion that seems to be represented by the Siberian I’ and South Chinese L4 excursion (see review at end of Stop 6A as well as discussion in Landing *et al.*, 1989, 2013b; Landing and Kouchinsky, 2016).

These small shelly fossil assemblages bracket the first recognized and named *Watsonella crosbyi* Zone (Landing *et al.*, 1989). The lithostratigraphy of Stop 6A characterizes the marginal Avalonian platform from SE New Brunswick, Cape Breton Island, the Burin Peninsula, and to North Wales in including a siliciclastic shelf succession that abruptly becomes wave dominated (middle–upper Chapel Island Formation in North American Avalon) that is abruptly overlain by a tide-dominated, sandstone-rich unit (Random Formation) (see Landing *et al.*, 2013a).

Stop 6A.1. Member/Lithofacies Association 3 of the Chapel Island Formation (75-135 m)

Description

Member/lithofacies association 3 consists of grey-green siltstone with fine laminae to thin beds of fine-grained sandstone with parallel lamination and ripple-scale cross-stratification. Elongate carbonate concretions form bedding-parallel nodule horizons. These horizons are

common and generally concentrated along thin sandstone beds. The upper bedding planes display current ripples, current lineations, and a limited range of trace fossils. Only unidirectionally oriented cross-laminae are preserved within the ripples, although some possible wave-ripple lamination exists near the top of member/lithofacies association 3. There is an upward decrease in sandstone percentage and an increase in the number of burrows noted within member/lithofacies association 3. The upper 15–20 m of this interval shows strong bioturbation that resulted in a loss of prominent bedding planes and internal sedimentary structures.

Interpretation

Paleocurrent data from this lithofacies indicate that sand was transported by unidirectional currents, at times in upper-plane-bed conditions as indicated by current lineations. Grading and sequences of sedimentary structures indicate deposition from waning flows below the influence of waves. The sandstone beds are thus interpreted as sub-wave-base, storm-generated deposits—namely distal tempestites. There is a gradational decrease in sandstone percentage and sandstone bed thickness from the more proximal sandstone and siltstone beds of member/lithofacies association 2, which show evidence of wave activity, and indicates a gradual deepening. The lack of instability features, so common in member/lithofacies association 2, is also likely due to the shift to deeper water, where the high sedimentation rates and high depositional slopes associated with the deltaic environment were not a factor.

Paleontology

Trace fossils are not very diverse, possibly in part because only upper bedding surfaces are exposed (no soles). *Helminthopsis* is visible on some bedding surfaces.

Bioturbation levels are considerably higher than in member/lithofacies association 2, as shown by the abundance of the mid-tier spreite trace fossil *Teichichnus rectus* (McIlroy and Logan, 1999; Mángano and Buatois, 2014).

Biom mineralized taxa (except for “*Ladatheca*” *cylindrical*) are absent in member/lithofacies association 3 at Little Dantzic Cone. However, *Watsonella crosbyi*, *Aldanella attleborensis*, and *Helcionella* sp. occur ca. 31 m below the top of lithofacies association 3 ca. 35 km to the NNE at the Fortune North section (Figure 2, locality FN). This diachronous lower base of the *W. crosbyi* Zone (39 m below the lowest occurrence of the same taxa at section 6A) reflects selective preservation (pyritization) of these aragonitic taxa in upper member/lithofacies association 3 at Fortune North (Landing *et al.*, 1989, 2013b).

Stop 6A.2. Lower Member/Lithofacies Association 4 of the Chapel Island Formation (135–185 m)

Description

The stop commences with the lowest of three stratigraphically successive limestone beds (LS 1, 2, and 3). These limestones are recognized in early reports on the section and display a progressive increase in bed thickness and diversity of sedimentary structures (Figure 34), as well as progressively more diverse biom mineralized taxa (Landing *et al.*, 1989). This lowest limestone bed caps a shoaling cycle of green to red mudstone. Above the bed is more green and red mudstone, as well as gray mudstone units. The limestone bed at this stop (LS 1) is a 15 cm thick, white micritic limestone that lacks any significant megascopic physical or biogenic structures. The lower 6–8 cm consists of partially coalesced nodules. Above, and filling in between the nodules, is a 2–3 cm-thick layer of red siliciclastic mudstone. The upper part of the bed is

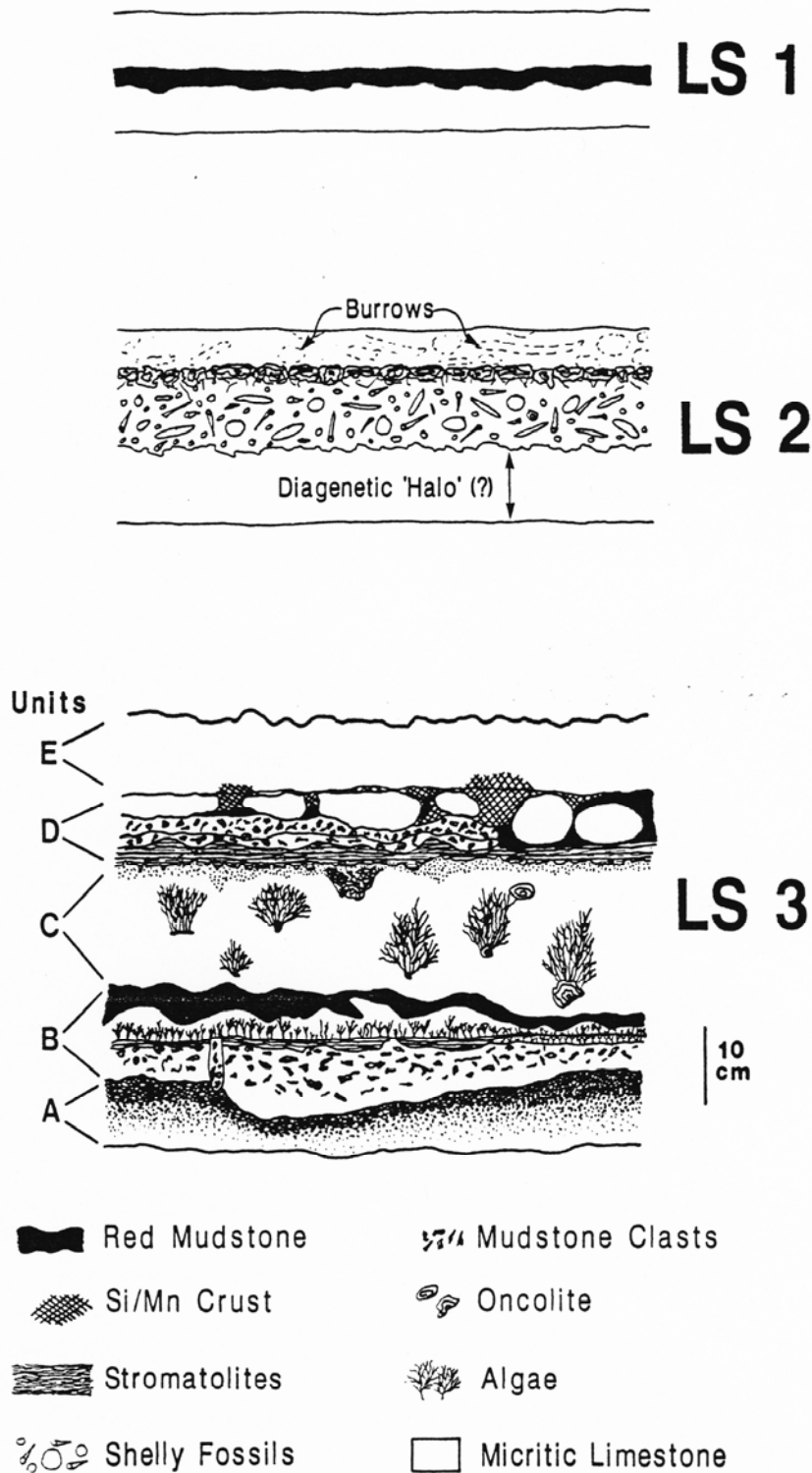


Figure 34. Sketches of limestone bed LS 1, LS 2, and LS 3 of member/lithofacies association 4 at Little Dantzic Cove. From Myrow and Landing (1992, fig. 4).

homogeneous white micritic limestone. A few specimens of the small leaf-like shelly fossil *Halkiera stonei* have been found in this bed (Landing *et al.*, 1989).

Above LS 1 is bioturbated purple, red, and green mudstone with laminae/thin beds of calcite-cemented sandstone. There is a general lack of exposed bedding planes and sedimentary structures. Carbonate concretions in the mudstone are internally massive and do not appear to have their stratigraphic position controlled by bedding. Centimeter-scale phosphate and pyrite nodules are not common but are locally abundant.

In comparison to the red mudstone, the green mudstone tends: (1) to be less bioturbated, (2) contains only thin laminae of sandstone, and (3) contains abundant carbonate nodules. The red and purple mudstone tends to contain more thin laminae to thin beds of sandstone and fewer and smaller calcite nodules. A few thicker sandstone beds (2.2–5.5 cm-thick) are found in the red mudstone and these contain parallel lamination, ripple cross-lamination, and ripple form sets.

Units of light to dark gray mudstone have abundant pyrite nodules, pyritic steinkerns (internal molds of small shelly fossils), and pyrite-replaced conchs, all of which are relatively rare in the green mudstone and absent in the red and purple mudstone. Small phosphate nodules (1–4 cm across) exist in this lithofacies, but are not abundant. The pyrite nodules are centimeter-sized, irregularly shaped, and have prominent orange-weathering halos. This facies is generally devoid of coarse-grained detritus (a few sandstone laminae), stratification, and calcite nodules.

Interpretation

Based on the upward transition from the below-storm-wave-base deposits of member/lithofacies association 3, and the intimate association with a shallow water limestone bed (see interpretations below for the other limestones and member/lithofacies association 4 in

general), the mudstone of member/lithofacies association 4 was deposited over a wide area of the inner shelf (below fair-weather wave base) in low-energy settings under conditions of relatively low sedimentation rate. A lack of sedimentary and biogenic structures in LS 1 suggests a quiet subtidal setting (deeper water than LS 2 and 3, described below) not colonized by algal mats or home to significant numbers of shelly fossils, as recorded in the overlying limestone beds.

The red and purple mudstone owes its color, lack of pyrite nodules, and paucity and small size of carbonate nodules to the early loss of organic matter under well-oxygenated diagenetic conditions. The presence of laminae and very thin sandstone beds indicates that bioturbation was inhibited at times. The green shale was deposited under slightly lower oxygen levels. The diagenetic conditions were not, however, sufficiently oxygen-depleted to generate abundant pyrite nodules or pyritic casts and molds of small shelly taxa, although a few horizons at Stop 6A.2 yield pyritized specimens of the *Watsonella crosbyi* Zone (Landing *et al.*, 1989; Figure 35).

Paleontology

Only locally in the bioturbated red and green mudstone are simple burrows preserved. Pyritized specimens of some of the earliest known shelly fossils are present in green and gray mudstone from 146–167 m. These include slender orthothecid/ “worm” shells ("*Ladatheca*" *cylindrica*) and a low diversity *Watsonella crosbyi* Zone fauna (Landing *et al.*, 1989).

Stop 6A.3. Second Limestone Bed (LS 2) in Member/Lithofacies Association 4 (185 m)

Description

This second limestone bed (LS 2) is exposed on a small promontory that may be difficult to reach along the shoreline from the north. It may be necessary to drop down from the trail

above. LS 2 (Figures 34, 35) contains abundant small shelly fossils. Some of the cone-shaped shells are remarkably large, with circular to slightly elliptical cross sections greater than 1 cm in diameter. The bed is dark red on fresh surfaces and averages approximately 24 cm in thickness. The lower 10 cm consists of structureless micrite with burrow mottles, but no obvious shelly fossils. Above this unit is 10 cm of “*Ladatheca*” and *Allatheca* conch-rich wackestone with irregular packstone lenses. The contact between these layers is very sharp, and irregular, displaying several centimeters or more of relief over very short distances.

Near the top of the bed is a surface characterized by a black, irregular, ropelike, anastomosing crust of silicified and iron and manganese oxide-mineralized siliciclastic mudstone. Stringers of this crust also parallel the cleavage, and penetrate a few centimeters downward into the limestone. The crust also surrounds low-relief, disc-shaped to irregular oncoids, 1–10 cm in diameter, that are scattered across this surface. The oncoids display an onion-skin concentric pattern of resistant limonitic red shale layers, with, in many cases, an outermost layer of mineralized Si-Mn crust. Above this oncolitic surface is 4–5 cm of nodular micritic limestone in shale.

Between LS 2 and LS 3 (185–220 m) is red mudstone. A few thin, burrowed, green calcareous siltstone beds are present in the red mudstone at 188.0 m and 188.12 m, with thicknesses of 2 cm and 1.5 cm, respectively. A few burrows in these beds show incomplete geopetal lime mud filling and a minor quantity of occluding white calcite cement.

Interpretation

LS 2 represents deposition in well oxygenated, moderate- to high-energy conditions (as shown by the oncoids). Subaerial exposure of the sediments during deposition cannot be

demonstrated conclusively. The abundance of oncoids may suggest a very shallow subtidal setting in which mat fragments, eroded from the intertidal area, were rolled around by tidal (?) currents or waves. The lower, unfossiliferous, micritic part of the bed may have formed by the deposition of lime mud or as a diagenetic “halo” below the overlying bed. In the first case, the irregular surface that separates the lower and upper parts of the bed would be interpreted as a submarine hardground surface, while the second alternative would have this contact represent a lower bedding plane from which a later, diagenetic layer grew downward likely as a caliche. Clear-cut evidence for hardground development (e.g., bored surfaces) is lacking. Well-defined pseudospar textures, typical of concretionary growth, are also absent.

The fossiliferous wackestone of LS 2 contains no evidence of significant winnowing or current reworking, as the orthothecid shells float in a matrix of lime mud. The oncolitic surface near the top of the bed may record maximum shoaling, while the wackestone below and the bioturbated mudstone may represent slightly deeper water conditions. Alternatively, the “halo” at the base of the limestone may be a caliche that shows subaerial conditions (see discussion in following text of caliche pseudospar under cap limestone (LS3) of member/lithofacies association 4).

Paleontology

A number of shelly animal taxa have their lowest appearance in this bed, indicating that the organisms are facies specific, and that their appearance records biofacies migration associated with shoaling (Landing *et al.*, 1989, 2013b, 2015c; Landing and Westrop, 2004; Figure 35).

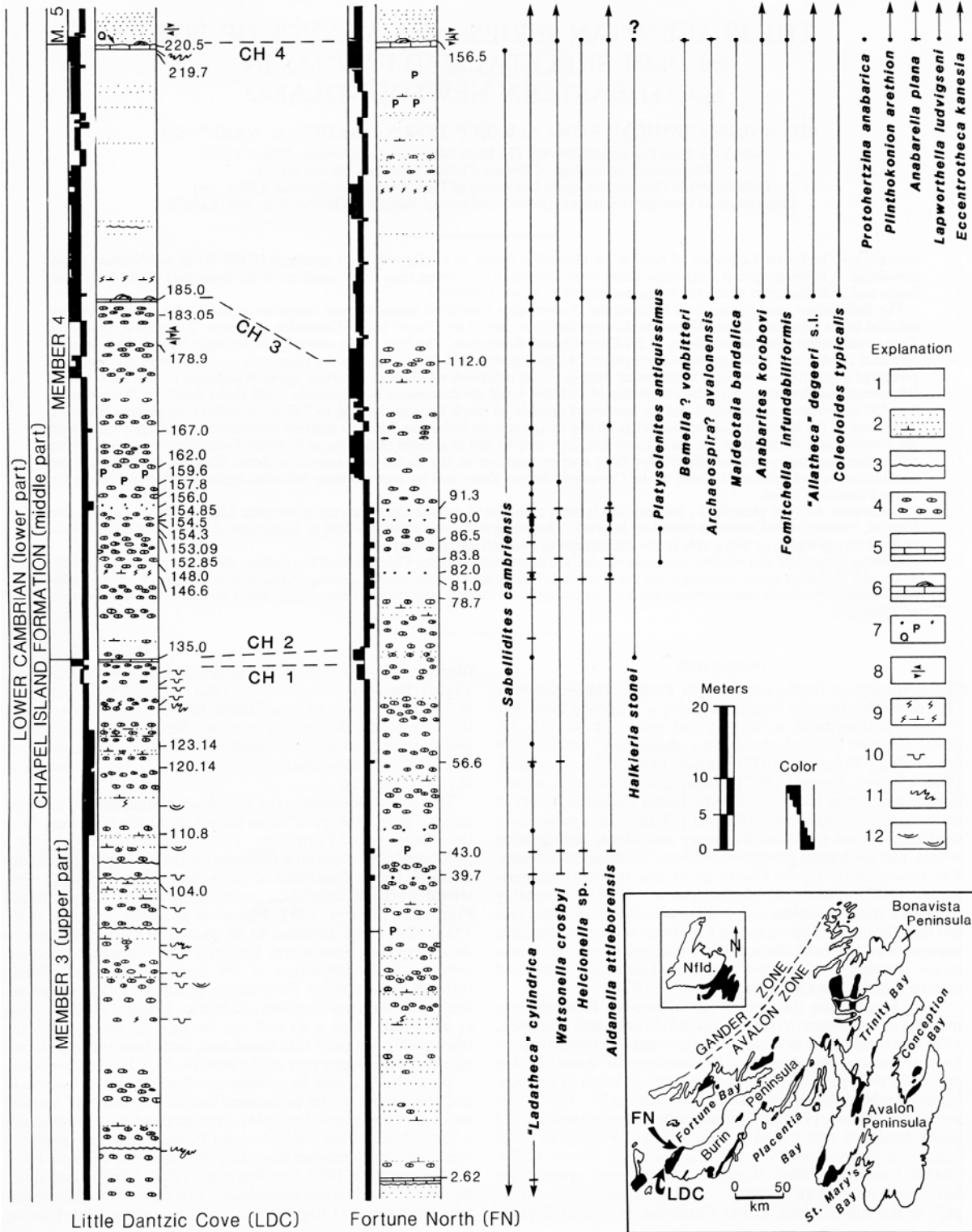


Figure 35. Thirty-nine meter differences in lowest occurrences ("FADs") of *Aldanella attleborensis* and *Watsonella crosbyi* (and a "Cambrian Stage 2" base defined by the "FAD" of either taxon) shown by correlation of middle Chapel Island Formation between Fortune North

and Little Dantzic Cove sections (Stop 6A; Figure 2, localities FN and LDC, and inset in figure). “CH,” correlation horizons, are nodular and bedded limestones used for correlation. Taxon occurrences at LDC shown by bold black dots; those at FN shown by horizontal dashes on teilzone line. Abbreviations: *l.a.*, lithofacies associations. “*Maldeotaia bandalica*” of figure now referred to *Fomitchella acinacinaformis* by Kouchinsky et al. (2017). Modified from Landing et al. (1989, fig. 1). The member/lithofacies association 4–5 contact (Avalonian Depositional Sequence 1–2 contact) just below top of thick limestone (ca. 156.4 m at FN; ca. 220.4 m at LDC). Explanation: 1, siliciclastic mudstone; 2, thin (mm- to 6 cm-thick) sandstone in lithology 1, locally calcareous; 3, ripples; 4, microsparite nodules (less than 5 x 10 cm in red mudstone, larger in green); 5, bedded nodular limestone (FN-2.62, LDC-135.0) and fossil hash wacke- to grainstone with lower caliche (LDC-185; -220; FN-156.5); 6, bedded limestones with oncoids or microbial build-ups; 7, pyrite (black dots), phosphatic (P) nodules, and Precambrian hydrothermal quartz, volcanic, and argillite clasts; 8, bedding plane shears/faults; 9, Planolites; 10, *Teichichnus* on bedding surfaces; 11, *Helminthopsis*; 12, *Teichichnus* in cross section. Colors: red (upper left); purplish-red; purple; purple with green streaks/plotches; green with purple streaks (center line); olive green; dark green; greenish grey (section LDC) and dark greenish grey (section FN); dark grey/black (lower right).

Stop 6A.4. Thick Limestone Bed (LS 3) at Top of Member/Lithofacies Association 4 (220 m)

Description

The third limestone unit (LS 3; Figure 34), which defines the upper limit of member/lithofacies association 4, is 50–80 cm thick. The unconformity surface that defines the Avalonian Depositional Sequence boundary 1–2 contact and the regional Quaco Road (lower)–Mystery Lake Member lies ca. 10 cm below the top of this bed (Landing 1996a; Figures 5, 7, 39). This composite limestone bed at Little Dantzic Cove, as well as LS 1 and 2, can all be precisely correlated 35 km to the NNE to the Fortune North section (Figure 2, locality FN; Figure 35, “CH horizons”). A few meters of red mudstone are just below the limestone bed, and a sharp transition above to the green siltstone and sandstone of member/lithofacies association 5

is disrupted by a fault that is present a few meters above the bed. Myrow *et al.* (1988) divide the bed into five divisions (Units a–e) (Figure 34), whereas Landing *et al.* (1989, fig. 3.4, units a–f) divided it into seven units. In the succeeding discussion, the limestone will be divided into five units.

At the base of the bed is a 38 cm pinkish unit (Unit a) of pseudospar calcite crystals within a red siliciclastic mud matrix (Figure 34). The pseudospar crystals grade upward in size from fine to very coarse sand. The textures of this unit indicate *in situ* displacive precipitation of pseudospar crystals within unlithified red mudstone matrix. Unit a could be regarded as a diagenetic “basal layer” or underbed (Eder, 1982; Meischner, 1967) that grew downward from the base of LS 3. Alternatively, unit a is best interpreted as an undifferentiated crystic plasmic fabric (Brewer, 1964) that reflects subaerial exposure and fluctuating ground water levels (see Landing *et al.*, 2013a, section 5c).

Unit b is a 7–15 cm thick stromatolitic interval. The base is an irregular erosion surface overlain by a chaotic zone containing angular to subrounded limestone intraclasts, oncoids, fragments of red mudstone (up to 1 cm in diameter), and well-rounded blue/black-weathering phosphatic shale clasts (up to 2 cm in diameter). Above, an extensive planar stromatolite layer is present and then a red siliciclastic mudstone layer, much of which is silicified and/or manganese-impregnated. The stromatolitic layers are broken by irregularly spaced, variably shaped pillars of sediment that locally extend below the level of the stromatolites into the chaotic material below and are interpreted as infillings of shrinkage cracks.

Unit c of the bed is a 5–25 cm-thick, massive, pink wackestone, locally containing finely disseminated quartz silt particles. There is local development of thin, dendritic structures interpreted to be rivulariacean algae (Figure 34). Widely dispersed at different levels are red and

green shale clasts up to 1 cm across and oncolites up to 2.5 cm in diameter, from which similar microbial forms project (Figure 34). Some of the oncolites and the shale clasts (green ones) have been silicified and Fe/Mn impregnated.

Unit d is similar to Unit b in that it contains conglomeratic material and stromatolitic layers. “Pillars” of sediment cut the stromatolites and either open downward to form tapered rooflike structures or widen and then narrow downward to form elongate bulbous shapes. Bedding-parallel views of these pillars reveal a distinct polygonal pattern, implying episodic desiccation. Within the stromatolites are sheet-crack structures and rare tepee structures, the former of which consist of thin, elongate seams of white sparry calcite. Red-weathering microbial structures, present along the base of these layers, represent some of the earliest known coelobiontic taxa (Myrow and Coniglio, 1991). An important feature, noted within the stromatolites of this unit, is a tepee structure 50 cm wide and 10 cm tall. There are also two mounded, lenticular bioherms of pink micritic limestone, one of which extends above the top of the bed. The large bioherm is 2 m wide at its base and 35 cm thick and extends 15–20 cm above the bed. The conglomeratic part of Unit d contains abundant red shale clasts, large limestone intraclasts, oncolites, well-rounded quartz pebbles (up to 1.5 cm in diameter), a variety of shelly fossils including abundant nested orthothecid hyoliths, and well-rounded volcanic clasts.

Unit e, overlying an irregular shale drape, consists of approximately 15 cm of pink/white fossiliferous wackestone (Figure 34). Isolated large blocks of silicified and partly silicified shale, some containing elongate bladed crystals of barite, are also present. Cut slabs show irregular surfaces with sharp overhanging walls that are interpreted as hardgrounds. The very top of Unit e has siltstone-lined depressions with siltstone (Figure 33d) and red shale clasts, and local accumulations of shell debris. Unit e has the most abundant *Watsonella crosbyi* Zone fauna (19

taxa) from the Chapel Island Formation at both the Little Dantzic Cove and Fortune North localities (Landing et al., 1989; Figure 35). This diverse fauna is regarded as a shallow-water fauna that reflects tempestite deposition with initial onlap of Avalonian Depositional Sequence 2 (Landing, 1996b; Landing and Westrop, 2004).

Interpretation

The features of LS 3 indicate deposition of a vertical succession of temporally changing microenvironments. The “micro-facies” include: (1) microbial mats, (2) microbially built (?) mud mounds, (3) low-energy, semirestricted (?) lime mudstone, (4) slightly higher-energy wackestone–packstone, (5) storm-sheet and channel coarse-grained debris, and (6) a basal subaerial caliche. These facies and the presence of desiccation (?) cracks, sheet cracks, and a tepee structure with relatively intact roof are consistent with deposition in low-energy intertidal settings. The tepee structure would correspond in character to Asserto and Kendall's (1977) “embryonic” stage of development of these features, which they consider diagnostic of lower intertidal settings.

The origin of the small fraction of quartz sand may have been either fluvial or eolian or both. Large volcanic clasts, rounded during fluvial transport, represent a thin scattering of terrigenous material (i.e., eroded Ediacaran basement) that was transported by streams flanking the tidal-flat region, possibly during riverine flood events.

Paleontology

This complex, amalgamated bed contains a high diversity assemblage of small shelly fossils (Figure 35), which, as in LS 2, resulted from shoaling and seaward migration of shallow

water environments with their indigenous, dominantly aragonite-secreting organisms (Landing *et al.*, 1989, Landing, 1992).

Member/Lithofacies Association 4: Summary Interpretation and Cambrian Stage 2 Correlation

The overall upward increase in red coloring of the mudstone facies within member/lithofacies association 4 parallels the progressively shallower water environments of deposition indicated by each successive limestone bed (LS1–LS3). Thus, member/lithofacies association 4 represents a large-scale upward-shoaling succession that records several smaller upward-shoaling cycles (parasequences), three of which are capped by limestone beds (Figures 6, 36).

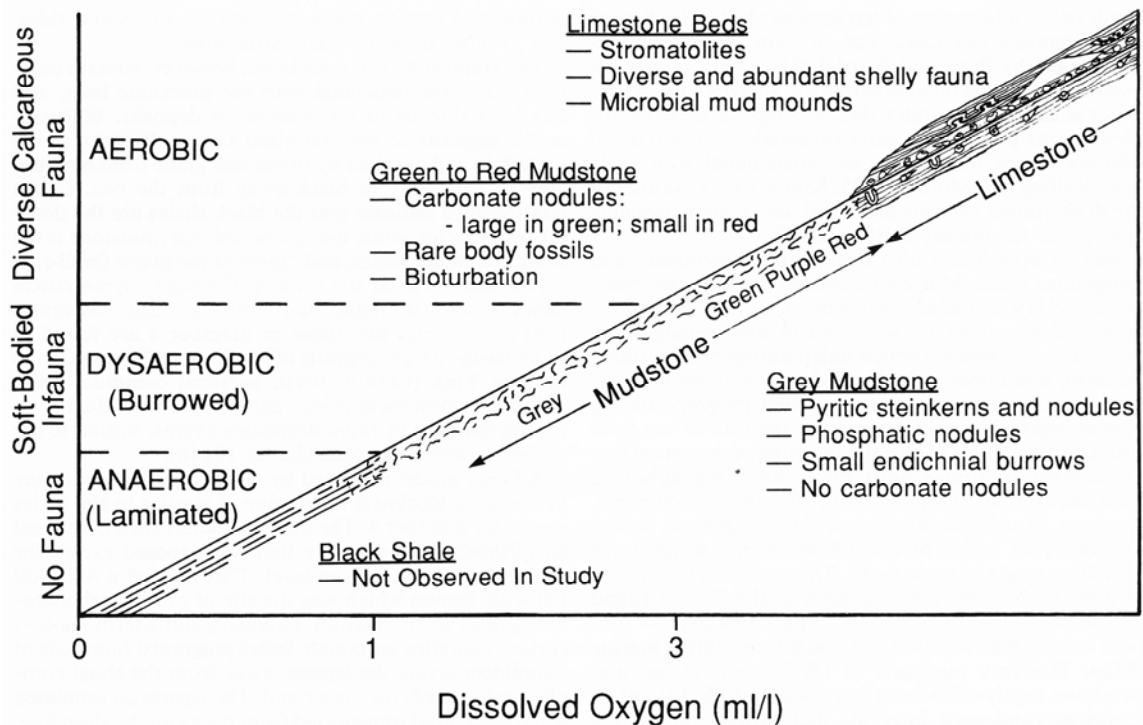


Figure 36. Biofacies model for oxygen-stratified basins. Levels of dissolved oxygen control the development and distribution of the three major biofacies. Sketch shows relationships present in member/lithofacies association 4. From Myrow and Landing (1992, fig. 8).

The color of the gray shale, and the presence of pyrite nodules and pyritic steinkerns, indicate low Eh conditions in the upper sediment column and possibly the bottom water that favored the preservation of organics (Figure 36). Pyritic steinkerns and pyritized conch replacement developed during early diagenesis in the upper few decimeters of sediment under low-Eh (dysaerobic) and low-pH conditions. Aragonitic conchs were dissolved and their molds filled with pyrite before the elimination of the molds by compaction in parts of the greenish mudstones at section LDC (Landing *et al.*, 1989; Myrow and Landing, 1992).

As shown by the diachronous lowest appearance of *Watsonella crosbyi* Zone shelly fossils (their lowest occurrence at Fortune North is 35 m below their lowest occurrence at Little Dantzic Cove; Figure 35), *W. crosbyi* Zone taxa and their shells may have been present and then dissolved with diagenesis. This would result because the redox conditions of the mud were locally not in the pyrite stability field, and the resulting conchs were neither filled nor replaced by pyrite. Thus, any evidence for the early occurrence of *W. crosbyi* Zone taxa was eliminated by non-pyritization and compaction to leave no record that a calcareous fauna was ever present in coeval strata at Little Dantzic Cove.

This evidence indicates that the lowest occurrences of *W. crosbyi* Zone taxa should not be used by themselves as a global standard for the base of Cambrian Stage 2. A more robust definition for a Stage 2/Laolinian Stage base is within the lower range of the *W. crosbyi* Zone and at the onset of the L4/P4/ZHUCE carbon isotope excursion (Landing *et al.*, 2013b; Landing and Kouchinsky, 2016).

The close association of the red mudstone with the peritidal limestone beds, and the fact that the gray mudstone exists near the base of the shoaling cycles and are never found in contact with red mudstone or with limestone beds, indicates that the gray mudstone is the deepest-water

deposit, and red mudstone the shallowest (Figure 36). Comparison of the Little Dantzic Cove with the Fortune North section approximately 35 km to the northeast provides support for this paleobathymetric distribution of lithofacies. Paleocurrent information indicates that this locality would have been more distal than Dantzic Cove, and indeed there is a greater abundance of gray mudstone although all three of the shallow water limestones persist from Little Dantzic Cove to Fortune North (Figure 35).

The data from member/lithofacies association 4 are consistent both with a basin stratified in terms of dissolved oxygen and with well-known biofacies models for modern oxygen-stratified shelves (Myrow and Landing, 1992; Figure 36). Such models have a dysaerobic facies (gray mudstone) with a low diversity assemblage of small infaunal burrowers. Bathymetrically, intermediate water depths occur between shallow nearshore aerobic facies (green to red mudstone) with abundant calcareous-secreting organisms and basinal low oxygen black muds that are largely devoid of organisms (Figure 36).

However, it should be noted that Avalonian dark gray–black shales were deposited in the range of wave-dominated deposition. They even show stromatolitic horizons as at the top of the Manuels River Formation (Landing and Westrop, 1998b, Highland Cove locality) and were not particularly deep. One of the best indicators of how shallow Avalonian black mudstone habitats were is the presence of conglomeratic caliches in black shale (e.g., Landing and Westrop, 2015).

Stop 6A.5. Middle of Member/Lithofacies Association 5 of the Chapel Island Formation (250–307 m)

Description

The strata include green to dark gray thin to medium bedded sandstone and siltstone (and minor laminated shale) that, due to minimal variation in grain size, weather without yielding

abundant sole or upper bedding surfaces. Sedimentary structures are commonly difficult to see due to this weathering pattern. Bed thickness increases upward within member/lithofacies association 5. The lower part consists of green-grey sandy micaceous siltstone with up to 6 cm-thick, very fine- and fine-grained sandstone beds. The sandstone beds have sharp lower surfaces and indistinct or highly disrupted upper surfaces, and the dominant internal sedimentary structures are parallel lamination and ripple cross lamination.

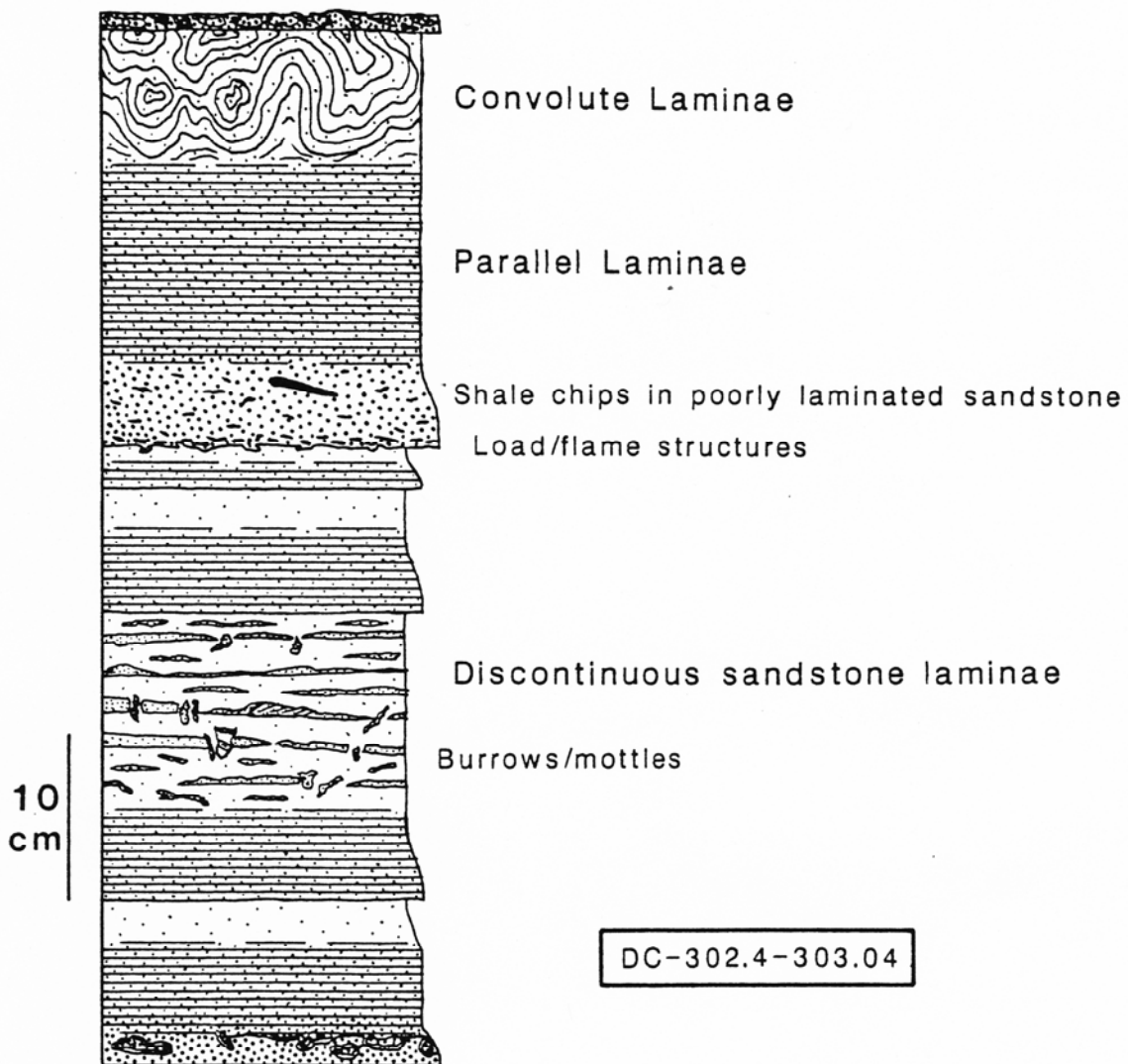


Figure 37. Detailed sketch of stratification in the middle of member/lithofacies association 5 at Little Dantzic Cove, Stop 6E.

These strata grade upward into siltstone and thin- to medium-bedded, gray, very fine-grained sandstone (Figure 37). Some sandstone beds have ball-and-pillow and flame structures and angular shale intraclasts. Parallel lamination is abundant and beds up to 25 cm thick are dominated by parallel laminae. There is a paucity of ripples and ripple cross-lamination in this facies. A few ripple forms are preserved. These are asymmetrical with unidirectional laminae.

Interpretation

The upward coarsening and thickening of sandstones indicate shoaling. The lack of evidence for wave reworking and upward transition into thick sandstone beds with HCS (next stop) suggests deposition of this part of member/lithofacies association 5 in sub-wave-base environments. Grading and sequence of structures in the beds are suggest deposition from decelerating sediment-laden turbulent flows, possibly as the distal hyperpycnal flows were distal equivalents of HCS beds. Flame and ball-and-pillow structures and the convolute bedding indicate rapid deposition of water-laden sand followed by liquefaction and loading.

These facies are interpreted as the deposit of the lower part of a prograding storm/wave-dominated system. Well established proximity trends for sandy storm-dominated shorelines (e.g., Brenchley *et al.*, 1979; Aigner and Reineck, 1982; Nelson, 1982) call for shoreward increase in bed thickness and a shift from sands dominated by ripple lamination to parallel lamination to HCS. These trends are seen vertically in this outcrop, indicating progradation.

Paleontology

Trace fossils are not common due to a lack of exposed soles and upper bedding surfaces. However, some discrete trace fossils are observed locally, including *Helmenthopsis abeli*,

Phycodes palmatum, *Psammichnites* isp., and *Teichichnus rectus* from the *Rusophycus avalonensis* Zone. Additional trace fossils, whose ranges extend from underlying ichnofossil zones, include *Paleophycus tubularis*, *Planolites montanus* and *Planolites beverlyensis*. Event beds are un- or sparsely bioturbated, but fair-weather deposits are intensely bioturbated.

Stop 6A.6. Middle–Upper Part of Member/Lithofacies Association 5 (310–317 m)

Description

The strata are transitional between the under- and overlying intervals. They consist of thin- to medium-bedded, parallel-laminated green sandstone and slightly calcareous, orange-weathering, bioturbated siltstone (Figure 38) with a few thin to medium HCS sandstones.

Interpretation

The presence of HCS beds indicates that shoaling progressed so that storm waves touched bottom. Deposition was above storm wave base but below fair-weather wave base.

Paleontology

Similar to the previous stop, the fair-weather siltstone is completely bioturbated, although the event deposits display a well-preserved primary fabric. Discrete, recognizable burrows include *Teichichnus rectus*, *Planolites montanus*, and *P. beverleyensis*.

Stop 6A.7. Upper Member/Lithofacies Association 5 (320–335 m)

Description

The strata in this interval consist of fine, micaceous sandstone of relatively uniform grain size and composition (Figure 38). This uniformity makes it difficult to see internal sedimentary

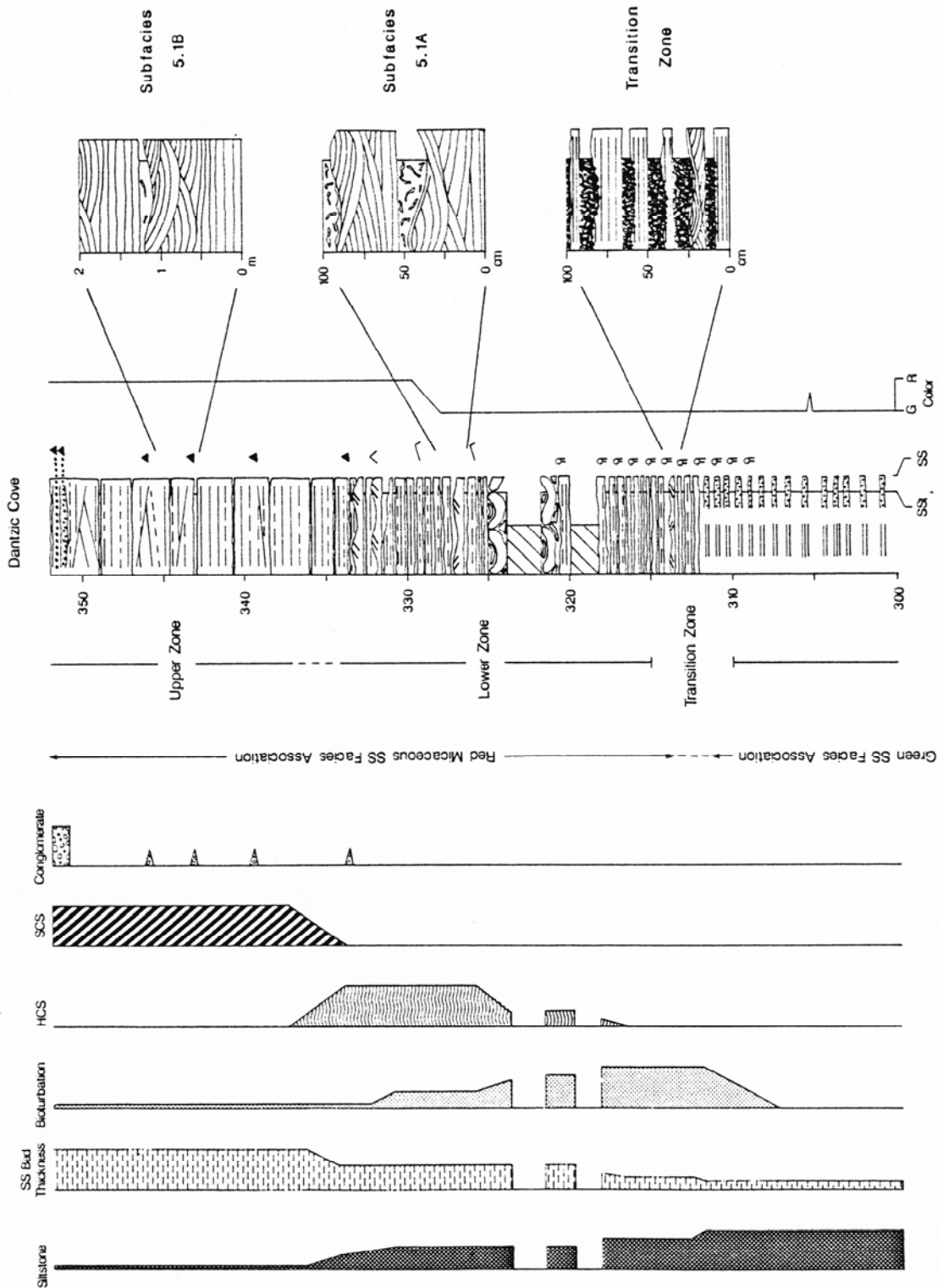


Figure 38. Features of section at top of upward-shoaling succession at top of member/lithofacies association 5 at Little Dantzic Cove.

structures in less-than-ideal lighting conditions. The sandstone ranges in color from greenish-pink to pink to deep red. This lithofacies is also present in the overlying Random Formation and was described by Hiscott (1982). The sandstone contains abundant parallel subhorizontal stratification separated by low-angle truncation surfaces, and well developed amalgamated hummocky cross-stratification. Large ball-and-pillow structures a meter or more across are present at the base of this interval.

Individual HCS beds are as thin as 10–20 cm, while the thicker beds are several tens of centimeters thick. Spacing of the hummocks ranges from 0.8 to 1.2 m. There are minor layers of shaley siltstone that drape hummocky surfaces. Their thickness is greater over swales than over hummocks. Bedding planes are locally covered with shale intraclast conglomerate, which are generally only a fraction of a millimeter thick. The long axes of these chips range from a few millimeters to 10 cm. Blue-weathering phosphatic shale chips are present in thin layers and lenses along internal truncation surfaces and floating within sandstone beds. Straight-crested wave ripples and high spacing-to-height ratios (vortex ripples) are on some surfaces oriented nearly at right angles to each other.

Interpretation

The stratification, dominated by planar lamination and HCS, represents deposition very near the HCS/upper-plane-bed boundary. The siltstone interbeds represent fair-weather suspension deposition. These formed during the waning stages of HCS deposition when flow decreased in intensity from that which produced the underlying HCS. The facies transitions, from parallel laminated beds below and swaley cross-stratified beds above (next stop), imply a

storm-influenced offshore-transition environment close to fair-weather wave base. The red color of these rocks reflects continued shoaling into well-oxygenated waters.

Paleontology

Tempestites in this interval show evidence of opportunistic colonization by a storm-related suite as reflected by the first occurrence of the *Skolithos* Ichnofacies in the Chapel Island Formation. In particular, the top of a sandstone bed at DC-331.9 shows *Diplocraterion* and other paired vertical burrows. This illustrates the appearance of a deep-tier suspension-feeding infauna by Cambrian Age 2 as a paleoecologic event decoupled from the explosion in ichnodisparity recorded by Fortunian ichnofaunas (Mángano and Buatois, 2014)

Stop 6A.8. Upper Member/Lithofacies Association 5 (335–352 m, 400–408 m)

Description

This stop is divided in half by a wide cobble beach that forms a covered interval. The lower part consists of massive beds of red micaceous sandstone that weather into smooth surfaces, with few pronounced bedding planes, due to uniformity in grain size (Figure 38). There are a few, thin, widely spaced siltstone layers that mark breaks between 1–2 m-thick sandstone beds. Phosphatic shale-chips make up conglomerate layers and also float within some sandstone beds. The sandstone has well-defined parallel laminae that are separated by low-angle truncation surfaces that locally take the form of isolated, wide, shallow scours filled with gently curved, low-angle (less than angle of repose) stratification. The scale and geometry of the scours and the stratification are similar to swaley cross-stratification (SCS; Leckie and Walker, 1982), which is thought to result from the amalgamation of HCS beds by the preferential loss of the topographic

highs (hummocks) and draping mudstones during erosion (Bourgeois, 1980; Dott and Bourgeois, 1982).

Granule conglomerate beds consist of a grain-supported mixture of pink and white quartz, and gray to black sedimentary rock fragments of shale, siltstone, and fine sandstone in a quartzitic sandy matrix. Quartz grains are generally well rounded and reach pebble size. Sedimentary rock fragments account for approximately 10% of the grains. The larger fragments, with long axes of tens of centimeters, tend to be more tabular. The conglomerate is present as beds 2–10 cm thick that cover low-angle erosional surfaces and broad, shallow scours. At 406.1 m, the conglomerate also forms large, symmetrical-crested, wave ripples with variable thickness (5–14 cm) and ripple spacings (32–45 cm).

Interpretation

The thickness of the sandstone beds indicates deposition in a shoreface setting. The thin siltstone interbeds reflect deposition in the lower shoreface. The red coloration of the rock supports oxygenated shallow-water conditions. A shoreface interpretation is supported by the fact that SCS-bearing facies are commonly transitional with underlying HCS beds and overlying beach facies (Hamblin and Walker, 1979; Bourgeois, 1980; Dott and Bourgeois, 1982; Leckie and Walker, 1982; McCrory and Walker, 1986). These strata contain more low-angle, subparallel and undulatory laminae and only a moderate percentage of recognizably isolated swales, which are typical of SCS (Walker, 1982). Transitions from lower, parallel-laminated divisions to upper, swaley-stratified divisions are common in this subfacies and are thought to have been generated as thick flows that decelerated from the upper-plane-bed stability field to a hummocky-bedform

field. The conglomerate beds were likely derived from local shoreline sources and swept offshore by storm surges.

Paleontology

The degree of bioturbation tends to be lower in this interval, as a reflection of the increase in hydrodynamic energy due to shallowing. A surface at 336.5 m contains *Psammichnites* sp. and *Protovirgularia* isp. Undetermined horizontal burrows occur at the top of member/lithofacies association 5.

Stop 6A.9. Random Formation

Description

The lower Random Formation is dominated by red micaceous sandstone identical to that in the upper part of member/lithofacies association 5 of the Chapel Island Formation (=upper Mystery Lake Member, Figure 4), but contain intervals of quartzarenitic sandstone and thin shale beds. This lower part of the formation represents a transition from storm- and wave-influenced deposition (red hummocky cross-stratified sandstone) to deposition under the influence of strong tidal currents (white quartzarenitic sandstone of the upper Random Formation).

Hiscott (1982) documented braided, erosional gully systems from two surfaces at this locality. The floors of these channels are littered with mudstone and sandstone intraclasts and phosphatic shale clasts. The sandstone mounds that are left as erosional remnants have abundant pothole structures that are also filled with clasts (Hiscott, 1982, fig. 9D). In addition to shale and mudstone clasts, large rip-up clasts of sandstone (up to 15 cm across) were noted on gully floors. These strengthen the conclusion that the sand was at least consolidated or perhaps semilithified

during erosion of the gullies. The shale and mudstone clasts that cover the gully surface locally exhibit vertical stacking. These clasts also line the bases of potholes, where they commonly are preferentially concentrated against one side of the pothole. These data indicate that prior to final deposition in these channels (and potholes), shale clasts were winnowed and reworked. Another interesting observation is that, aside from the prominent parallelism of the gullies oriented at $\sim 045^\circ$, a secondary erosional pattern cuts across the outcrop at $\sim 145^\circ$. Many of the potholes cut into the sandstone ridges are elongated parallel to this secondary direction. The flows that cut the prominent channels record the transport of sand, as well as mudstone and sandstone intraclasts and shale chips, into deeper water. The secondary channel pattern may represent modification of the sea floor by shore-parallel currents.

Paleontology

Crimes and Anderson (1985) describe a wide variety of complex trace fossils from the Random Formation. The trace fossils that they noted from this locality include a wide variety of vertical dwelling burrows from the *Skolithos* Ichnofacies. These include *Arenicolites* isp., *Diplocraterion* isp., and *Skolithos* isp., as well as such horizontal trails and burrows from the *Cruziana* Ichnofacies as *Helminthopsis abeli*, and *Planolites beverleyensis*.

Stratigraphy and Biomineralized Biotas of the Chapel Island Formation

The oldest diverse skeletal fossils in SE Newfoundland were first illustrated from the Little Dantzic Cove (LDC) section (Figure 2, locality LDC; Figures 35, 39) (Bengtson and Fletcher, 1983). Section LDC and the Fortune North (FN) section faunas have been described (Landing *et al.*, 1989; McIlroy and Szaniawski, 2000; Figure 35) and a preliminary $\delta^{13}\text{C}_{\text{carb}}$

record (Brasier *et al.*, 1992; Figure 40) has been documented. These data have led to discussion of LDC and the laterally equivalent section at Fortune North (Figure 2, locality FN; Landing *et al.*, 1989) as important in definition and correlation of informal Cambrian Stage 2, which has

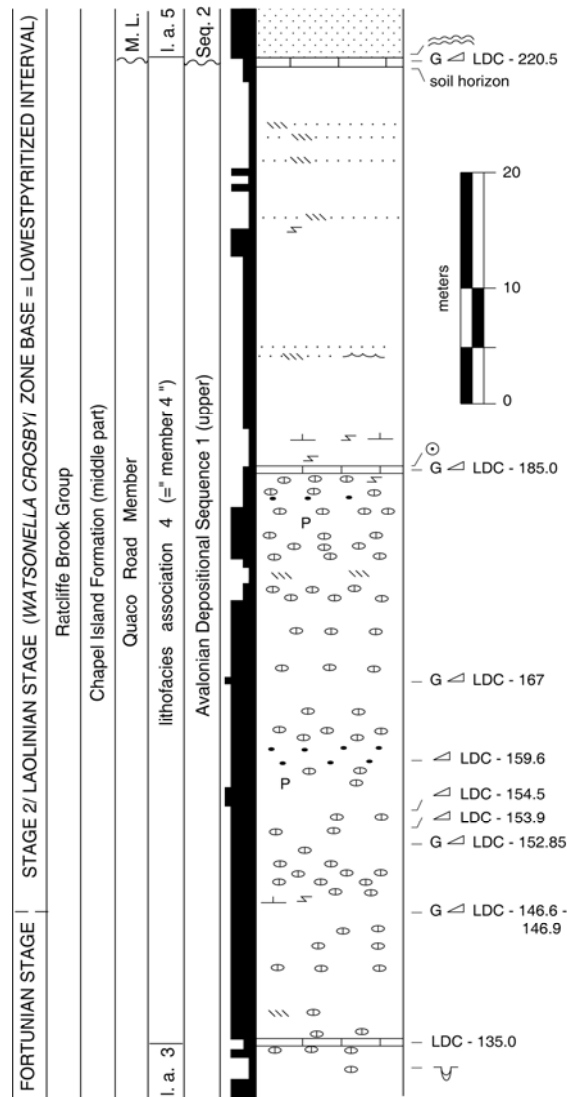


Figure 39. Generalized section of the Chapel Island Formation at Little Dantzic Cove (Stop 6A, compare with Figure 35). Abbreviations: *l. a.*, lithofacies association, *M. L.*, Mystery Lake Member; *Seq.*, Avalonian depositional sequence. Sample numbers give meters above base of measured section (e.g., LDC-135.0) Explanation of lithologic symbols in Figure 8. Modified from Landing and Westrop (1998a, fig. 12).

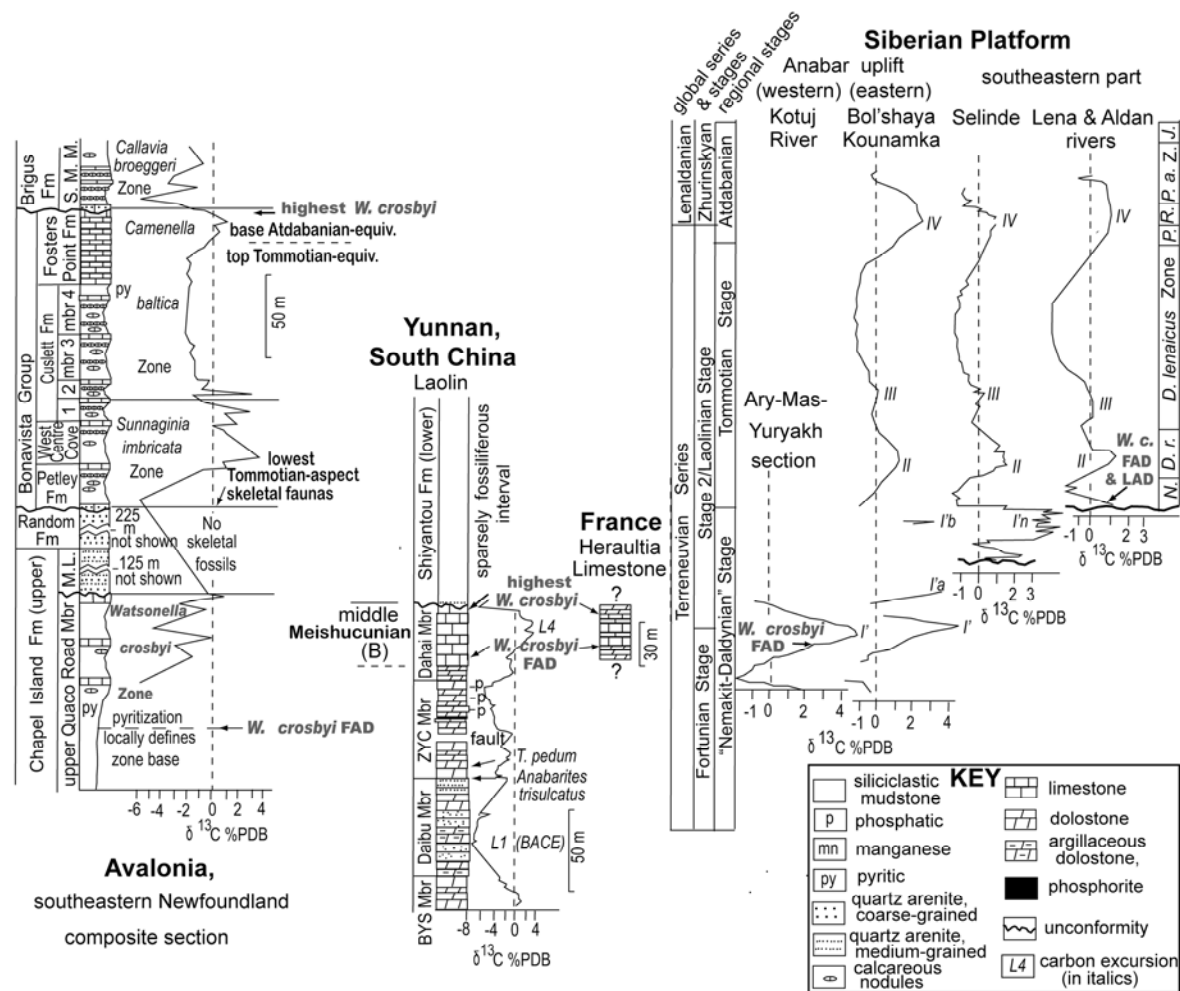


Figure 40. Lower Cambrian carbon isotope–faunal correlations. Notes: Avalonian carbon isotope stratigraphy (Brasier et al., 1992) and bio- and lithostratigraphy (Landing et al., 2013b, and references therein); Fosters Point Formation thickness increased four-fold for figure; carbon isotope stratigraphy of Lena–Aldan river region (Kirschvink et al., 1991; Brasier et al., 1994a, b), Bol’shaya Kuonamka (Kouchinsky et al., 2001), and Selinde (Kouchinsky et al., 2005); coeval Siberian and Avalonian stratigraphic units and isotope excursions aligned; South China Platform succession (Li et al., 2009). Abbreviations: Fm, Formation; Mbr, Member; M.L., Mystery Lake Member; T., Trichophycus; W., Watsonella. Siberian zones: N., Nochorocyathus sunnaginicus; D. r., Dokidocyathus regularis; D. lenaicus, Dokidocyathus lenaicus; P., Profallotaspis jakutensis; R., Repinaella; D. a., Delgadella anabara; J., Judomia. Correlation of Atdabanian Stage base into upper Fosters Point Formation based on carbon isotope data in Brasier et al. (1992), although faunas with Rhombocorniculum cancellatum in upper Cuslett Formation also suggest an Atdabanian-correlation. Modified from Landing et al. (2013b, fig. 5)

with addition of Heraultia and Ary-Mas-Yuryuakh sections and deletion of several Siberian and Chinese sections. Laolinian Stage, Lenaldanian Series, and Zhurinskyan Stage proposals (Landing et al., 2013b), submitted to Cambrian Subcommittee in 2013, follow Remane et al. (1996) guidelines.

also been designated as the “Laolinian Stage” of the upper Terreneuvian Series (Landing *et al.*, 2013b; Landing and Kouchinsky, 2015).

Section LDC shows the wave-dominated deposits of the upper Chapel Island Formation—ca. 220+ m of green siltstone (member/lithofacies association 3), green and red mudstone with scattered calcareous nodules and several limestone beds (member/lithofacies association 4), and ca. 150 m upward coarsening succession of mostly fine-grained sandstone of member 5 (Figures 5, 35). The formation is overlain by feldspathic and quartz arenite tidal deposits of the Random Formation (Hiscott, 1981).

Avalonian Depositional Sequence 1–2 Boundary

A sequence boundary lies almost at the top of the complex limestone bed at ca. 220 m that marks the top of member/lithofacies association 4 (Bengtson and Fletcher, 1983; Narbonne *et al.*, 1987; Landing *et al.*, 1989; Myrow *et al.*, 2012) at the LDC and Fortune North sections (Figure 35, 30).

However, the very top of this limestone bed (ca. 10 cm of purple “*Ladatheca*”–”*Allatheca*” packstone) is the base of Avalonian Depositional Sequence 2 (see Landing, 1996a; Landing and Westrop, 1998a; Rees *et al.*, 2014). As detailed above, this tempestite limestone overlies a highstand limestone that consists vertically of (1) red mudstone, (2) a light pinkish-white oncolitic fossil packstone, (3) thin fossil hash packstone that is piped down on fractures

into the middle of the bed (nodular pink fossil wackestone), and (4) several cm of dark-green, Fe/Mn oxide-impregnated cryptomicrobial laminate (Landing *et al.*, 1989, fig. 3.4). Coarse calcite crystals impregnate the two lower parts of the beds. They form an alteration zone with undifferentiated crystic plasmic structure and subhorizontal displacive fractures that indicate subaerial conditions and fluctuating ground-water levels (e.g., Sedov *et al.*, 2008).

Litho- and Biofacies Linkages of “Stage 2” Faunas

Members/lithofacies associations 3 and 4 record progressive upward shoaling (Landing *et al.*, 1989; Myrow and Landing, 1992; Myrow *et al.*, 2012). The “*Ladatheca*” *cylindrica* Zone (with *Sabellidites cambriensis*) features eurytopic taxa and exists in dysoxic olive-green mudstone at sections LDC and FN (Figure 35; Landing and Westrop, 2004).

The base of *Watsonella crosbyi* Zone faunas at Fortune North (*W. crosbyi*, *Aldanella attleborensis*, and a helcionelloid) records preservational windows in dark green mudstone. This dysoxic facies allowed pyritization of conchs well below “correlation horizon 1” at Fortune North (Figure 35). Some of the *Watsonella* conchs and “*Ladatheca*” tubes with opercula in place are *in situ* (vertically embedded; Landing, 1989, 1993). Similar conch pyritization appears only 39 m higher and above “correlation horizon 2” within member 4 at section LDC (Figure 35).

Watsonella crosbyi Zone faunas become increasingly diverse with shoaling in the uppermost member’ lithofacies association 4 at both sections LDC and FN. This suggests the origin and diversification of the oldest skeletalized faunas in peritidal environments subject to intense UV-B radiation and desiccation (Landing and Westrop, 2004; Meert *et al.*, 2016). The presence of *W. crosbyi* Zone faunas above and below the member 4–5 (Avalonian Depositional Sequence 1–2) unconformity suggests that the hiatus was not very long.

Limitations to Cambrian Chronostratigraphy of Simple Fossil FADs

Eastern Newfoundland shows that the FADs of *Watsonella crosbyi* and *Aldanella attleborensis* are inadequate by themselves to define a base of a global Cambrian Stage 2 (Landing *et al.*, 2013b; Landing and Kouchinsky, 2015). The lowest occurrences of both taxa are 39 m lower than at section LDC at the nearby Fortune North section as this more distal section was poorly oxygenated and *W. crosbyi* Zone assemblages occur lower in pyritiferous mudstones.

Preliminary work on carbon isotope stratigraphy (Figure 40) shows that progressively more strongly positive $\delta^{13}\text{C}_{\text{carb}}$ values occur through member 4 at LDC (Brasier *et al.*, 1992). This carbon isotope excursion suggests the I' and L4 excursions in north-central Siberia and South China (Landing *et al.*, 2013b; Landing and Kouchinsky, 2015), and supports correlation of the LDC and FN faunas with the sub-Tommotian upper “Nemakit-Daldynian” Stage of Siberia and middle Meishucunian Stage of South China originally suggested by Landing *et al.* (1989) and Landing (1994).

On the walk back south to Little Dantzic Cove Brook, the wave-dominated sandstones of the Mystery Lake Member of the Chapel Island Formation (ca. 150 m) and overlying feldspathic quartz arenites of the tidalite Random Formation (175 m) will be examined (Myrow *et al.*, 1988, 2012).

Stop 6B. Little Dantzic Cove Brook (June 26; EL, SRW, PM)

Generalized Stratigraphy and Significance

The upper Random Formation immediately north of the outlet of Little Dantzic Cove Brook (Figure 2, locality LDCB; Figure 41) includes a 30–50 cm-thick bed of quartzite boulder

conglomerate with phosphatic shale, vein quartz, siltstone, and chert(?) pebbles. This bed is overlain by 4.3 m of fissile sandstone with <50 cm-thick trough cross-beds and an uppermost 1.0 m of micaceous, purplish–bleached coarse-grained sandstone (paleoflow to 70°) (Myrow *et al.*, 1988).

A major unconformity is marked by the base of the Brigus Formation—a ca. 30 cm-thick calcareous quartz arenite with limonitic, low domal stromatolites. This bed is overlain by a 1.2 m-thick interval with three beds of massive, condensed limestone separated by thin red mudstone with calcareous nodules (Figure 41). The Bonavista Group (upper Cambrian Stage 2–lower Stage 4) is absent as on other Avalonian marginal platform successions in North American and British Avalonia (Figures 3, 5). Thus, the Random–St. Mary’s Member contact at Stop 6B represents the Avalonian Depositional Sequence 2–4A unconformity (Landing, 1996a; Figures 3, 5). The basal limestones of the Brigus Formation represent a highly condensed sequence with numerous hardgrounds with embayed and overhanging surfaces.

Callavia broeggeri Zone trilobites occur in these lowest limestones (Hutchinson, 1962) but the rarity of their broken sclerites reflects the proximal, peritidal environment of this onlap facies (Landing and Westrop, 2004). *Callavia broeggeri* Zone trilobites are abundant in more distal calcareous nodules in red siliciclastic mudstones through the thin 12.6 m-thick Brigus Formation at LDCB (only the St. Mary’s Member is represented). Wave cross-laminations and *Collenia*-type stromatolites at 6.5 m show relatively shallow-water deposition of the upper red mudstones of the Brigus at LDCB. It should be noted that the claim that the Avalonian Cambrian succession represents deeper-water deposition (Álvaro *et al.*, 2013) is unfounded (compare Landing *et al.*, 1989; Myrow and Landing, 1992). Indeed, even Avalonian dysoxic black

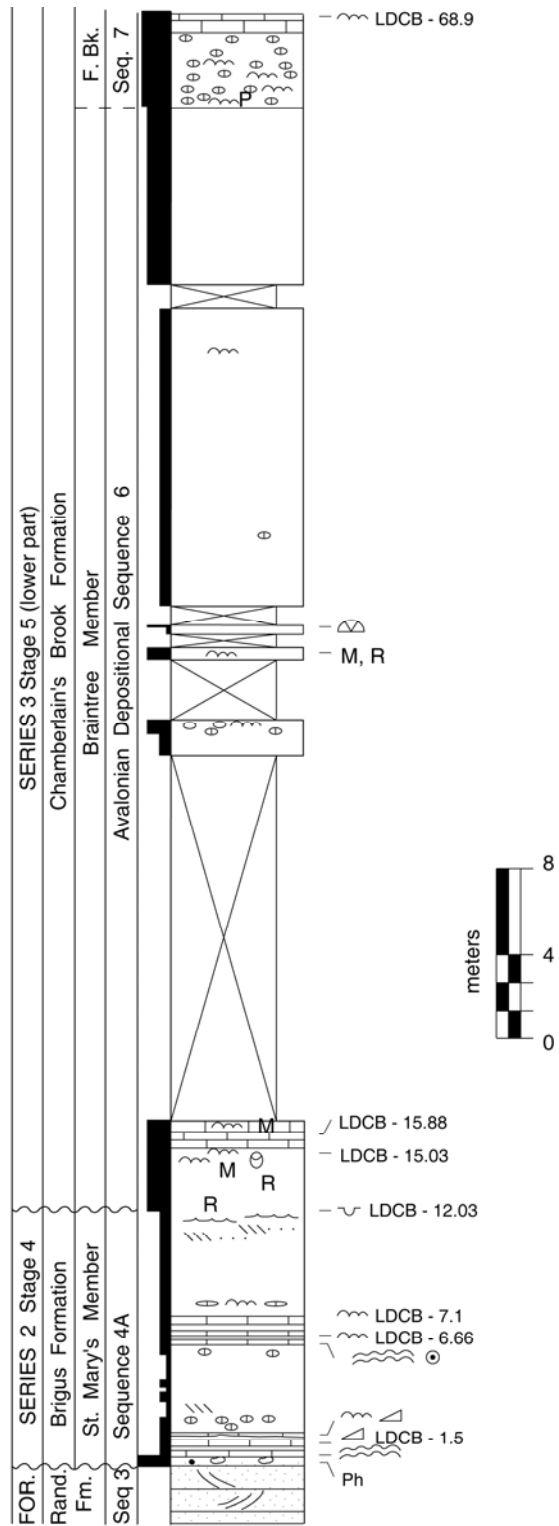


Figure 41. Little Dantzic Cove Brook succession (Stop 6B). Abbreviations: Fos. Bk., Fossil Brook Member. Lithologic symbols explained in Figure 8. Modified from Westrop and Landing (1998a, fig. 13).

mudstones show intertidal stromatolites (Landing and Westrop, 1998a, fig. 10) and subaerial caliche horizons (Landing and Westrop, 2015)].

The base of the Chamberlain's Brook Formation (12.6 m above Random) is marked by the abrupt appearance of brown-gray, silty mudstone with rhodochrosite nodules. The presence of *Eccaparadoxides bennetti* at 15.1 m suggests that the hiatus between the Brigus Formation (St. Mary's Member) and lower Chamberlain's Brook is comparable to that at Red Bridge Road (Stop 1) and represents the Avalonian Depositional Sequence 4A–6 unconformity (Figures 3, 5). Red and green-grey mudstones with condensed manganiferous limestones of the Braintree Member compose most of the Chamberlain's Brook Formation (12.6–66 m). These strata are sharply overlain by dark gray mudstones with nodules and a bedded limestone of the Fossil Brook Member (Avalonian Depositional Sequence 7) at the top of the of the Chamberlain's Brook (e.g., Hutchinson, 1962; Landing and Westrop, 1998a; 66–75 m) in the core of the LDCB syncline.

Stop 7. Northern Branch Cove: Lower–Middle Cambrian (Stages 4–5) Boundary Interval (June 27; EL, SRW, GG)

Location

The traditional Avalonian lower–middle Cambrian boundary interval at the Brigus–Chamberlain's Brook formational contact is exposed in low sea cliffs on the east shore of St. Mary's Bay (Figure 2, locality BC). Drive Rte. 100 north then east of Branch village. Park several hundred meters south of the wooded, low hill crest. Walk east through pastures to the cliff edge and descend a (very!) steep path to the shore cliffs.

Generalized Stratigraphy and Significance

The upper lower–lower middle Cambrian succession at Branch Cove (Figure 42) is located ca. 2 km NE of Branch village. This particularly thick Brigus–Chamberlain’s Brook interval in the SW St. Mary’s Peninsula accumulated on the depocenter of the St. Mary’s-east Trinity axis (Figure 2).

Stop 7 shows red siliciclastic mudstones with a brownish cast of the upper Jigging Cove Member of the Brigus Formation, the traditional upper lower Cambrian unit of Avalonian SE Newfoundland, and overlying manganiferous green mudstones of the traditional lower middle Cambrian Chamberlains Brook. The contact of the formations is at 103.05 m in the measured section.

Irregular, lensing, large green “splotches” of green mudstone with purple margins are common in the slightly calcareous Jigging Cove Member mudstones. Loose blocks of these green mudstones yield the typical West Gondwanan trilobite *Hamatolenus* (identified as *H. sp. aff. maroccanus* by Fletcher, 2006), which occurs with the earliest paradoxidids, and thus with traditional middle Cambrian trilobites in both Morocco and SE Newfoundland (Geyer and Landing, 2004).

Small calcareous nodules with the lowest middle Cambrian (lower Amgan Stage of Siberia) trilobite *Ovatoryctocara granulata* are reported about a meter below the contact with the Chamberlain’s Brook Formation (Fletcher, 1972, 2003, 2006). Repeated attempts to find this nodule horizon and its trilobites have failed (Landing *et al.*, 2013b), and it is possible that the few recovered specimens came from scattered, now completely exploited, small carbonate nodules at a single horizon. Volcanic ashes occur through the lower Jigging Cove at Branch

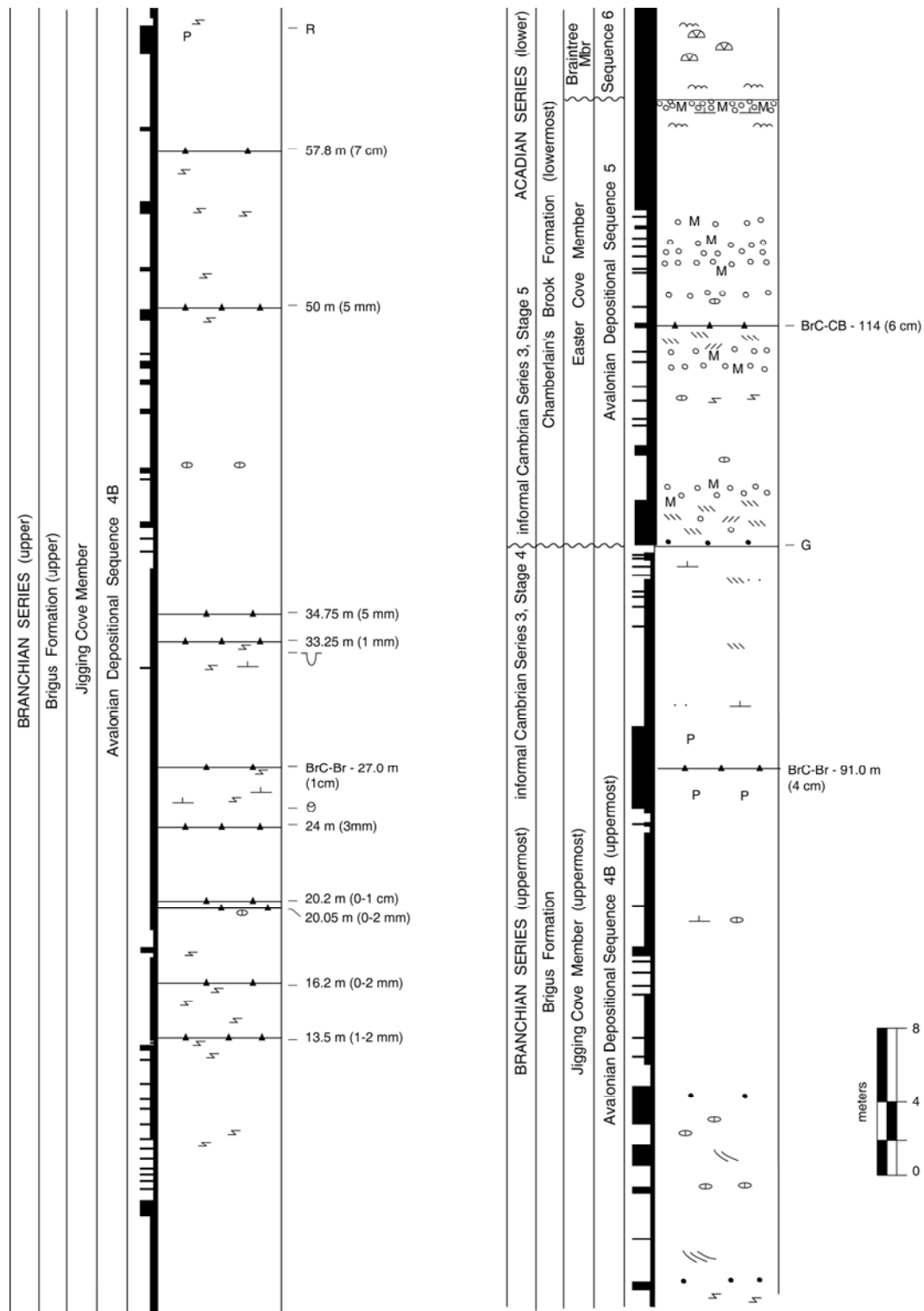


Figure 42. Terminal lower Cambrian–lower middle Cambrian boundary interval on north side of Branch Cove, western St. Mary’s Bay (Stop 7). Explanation of lithologic symbols in Figure 8. Modified from Landing and Westrop (1998a, fig. 14).

Cove, with the highest ash at 91 m (EL, in preparation) and not far below the formational contact.

Fletcher (1972, 2003) proposed a highly detailed trilobite zonation for the upper Jigging Cove Member (his “Branch Cove Member”) but subsequently (Fletcher, 2006) referred the interval to an undivided *Cephalopyge notabilis* Zone with the eponymous species (now *Moroccanus notabilis*), *Condylopyge eli*, *Serrodiscus occipitalis*, *Acidiscus taconicus*, *Hamatolenus aff. marocanus*, *Strenuaeva nefanda*, and unidentified paradoxidid fragments, among others.

Unconformable Brigus–Chamberlain’s Brook, Avalonian Depositional Sequence 4B–5, Lower–Middle Cambrian Contacts

The Brigus–Chamberlain’s Brook formational contact has repeatedly been termed “conformable” at all contacts (Fletcher, 1972, 2003, 2006; Fletcher and Brückner, 1974). However, examination of the contact interval at Branch Cove shows that the top 15 cm of the Jigging Cove Member is bleached to a light brown color not found elsewhere in the upper Brigus, while the base of the Chamberlain’s Brook is a horizon with large, scattered quartz grains; phosphate granules; and wave-sorted, sparse, manganese nodules. These relations are consistent with subaerial exposure and marine transgression at a depositional sequence boundary (Avalonian Depositional Sequence 4B–5 contact; Landing, 1996a; Landing and Westrop, 1998a, b).

The unconformity likely marks a relatively short hiatus as a number of upper Brigus taxa (e.g., *Moroccanus notabilis*, *Condylopyge eli*, *Eoagnostus rodnyi*) persist from the Brigus into the lower Chamberlain’s Brook Formation, where they are accompanied by traditional middle

Cambrian paradoxidid trilobites (*Acadoparadoxides harlani*) with *Kiskinella cristata* (Fletcher, 2006). This *K. cristata* Zone persists through the lower 27 m of the Chamberlain's Brook. Landing (1996a) had termed this 27 m-thick interval an "unnamed member." A likely sequence boundary at its top (Avalonian Depositional Sequence 6–7 unconformity) is marked by a 90 cm-thick, calcareous, packed manganese nodule bed with erosion recorded at the tops of the uppermost nodules. Fletcher's (2006) designation "Easter Cove Member" is appropriate for this unit that is preserved only on the southern St. Mary's-east Trinity axis.

Stop 8. Jigging Cove: Avalonian Depositional Sequence Unconformities on the Inner Platform (June 27; EL, SRW)

Location

The Jigging Cove section, located ca. 10 km NE of Branch village and on the NW shore of Jigging Cove (Figure 2, locality JC), is a completely exposed, E-dipping, lower Cambrian section at the outlet of Red Head River. Travel Rte. 100 further north from the Branch Cove section to the bridge over Red Head River and park. Walk the dirt road east and downstream on the south side of the river; cross the river (you will get wet feet), and examine the section from the top of the white Random Formation (Figure 43).

Generalized Stratigraphy and Significance

The Jigging Cove section lies just E of the axis of the St. Mary's-east Bonavista (syndepositional) axis and has a very thick section in the Brigus–Manuels River formation at Jigging Cove Head (Hutchinson, 1962, section 13). A continuous section in this inner platform

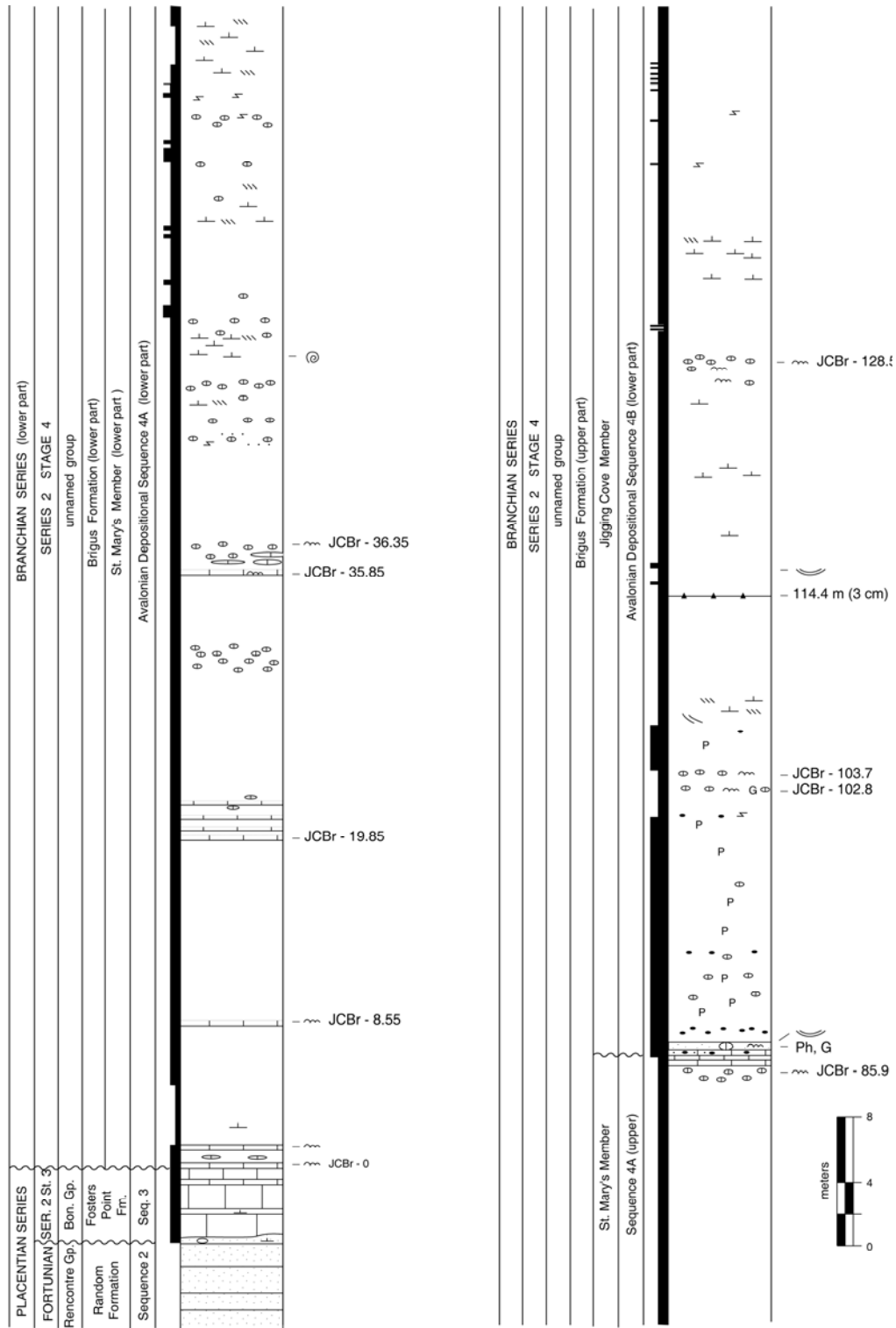


Figure 43. Lower lower–upper lower Cambrian (Fortunian–upper Stage 4/upper Lenaldanian Series) at Jigging Cove, western St. Mary’s Bay (Stop 7), with type sections of the St. Mary’s and

Jigging Cove members of the Brigus Formation (Landing and Westrop, 1998b). Explanation of symbols in Figure 8. Modified from Landing and Westrop (1998a, fig. 15).

succession does not include older terminal Ediacaran–lowest Cambrian units (Rencontre and Chapel Island Formations) characteristic of the marginal platform and the Random Formation unconformably overlies the middle Ediacaran, upper Musgravetown Group (Fletcher, 2006, Map 2006-02).

In addition, almost all of the Bonavista Group (which reaches 275+ m in the Placentia-Bonavista Axis to the east; Figures 2, 5) is absent. Only the Fosters Point Formation (4.5 m; uppermost *Camenella baltica* Zone) at the top of the Bonavista Group is present, and unconformably overlies the Random Formation (Figure 43). This unconformity and its duration is identical to the Jeas Member (quartzite) and overlying Home Farm Member (limestone) unconformity within the Hartshill Formation at Nuneaton, Warwickshire (Brasier *et al.*, 1978; Brasier and Hewitt, 1979; Brasier, 1984, 1989; Landing, 1996; Landing *et al.*, 2013b). The unconformities at Stop 7 and in Warwickshire represent the Depositional Sequence 2–upper 3 unconformity on the Avalonia paleocontinent.

The base of the Fosters Point Formation is a calcareous, limonitic reworked quartz arenite bed (35–50 cm-thick) with shale clasts. *Coleoloides*-dominated wackestone dominates the overlying 3.0 m of the Fosters Point. The regionally extensive, *in situ* *Coleoloides*-mud mound biostrome lies near the top of the Fosters Point (3.4–3.55 m; also Stop 10, Smith Point, June 28).

The top of the Fosters Point (0.95 m-thick) is a small shelly fossil and hyolith wackepackstone bed (with Fe-Mn oncoids and low domal stromatolites) truncated at the Avalonian 3–4A depositional sequence boundary (Figures 5, 45). The very base of the Brigus Formation is a

15 cm-thick, pinkish, trilobite wacke–packstone bed with a *Callavia broeggeri* Zone fauna. Fletcher (2006, p. 45) acknowledged this sequence boundary within his “Smith Point Formation,” without noting the existing Avalonian depositional sequence succession and revised stratigraphic nomenclature (Landing, 1996a, 2004; Landing and Westrop, 1998b). Fletcher assigned this lowest *Callavia*-bearing limestone bed variously to an undefined “Clifton Limestone Member” (with unspecified type section, thickness, and lithology; e.g., Fletcher, 2003) and later to a “Broad Cove Member” (Fletcher, 2006). However, it is simply a relatively thin condensed limestone at the base of the Brigus Formation. Trilobite-bearing limestone at the base of the Brigus can be locally absent, with red mudstone forming the Brigus’ base, or the basal Brigus limestone can range to 3.4 m-thick in SE Newfoundland (Landing and Benus, 1988b).

The lower ca. 85 m of the Brigus Formation at Stop 8 (Avalonian depositional sequence 4A, type section of St. Mary’s Member; Landing and Westrop, 1998a, b) crops out along the shore to the north along Jigging Cove as largely structureless (burrow-churned), red and purple mudstone with greenish lenses. Thin nodular to amalgamated nodular limestones often developed on large horizontal traces occur through the St. Mary’s Member.

The top of the St. Mary’s Member at Jigging Cove includes trilobite-bearing lime nodules in red mudstone (84.7–86.0 m) overlain by an exceptionally thick (in terms of Avalonian limestones), highstand trilobite wacke–packstone (86.0–86.8 m; with the highest *Callavia broeggeri*, *Serrodiscus bellimarginatus*, and *Strenuella strenua* specimens). The top of this bed has an undulatory (eroded top) with limonitic oncoids and small domal stromatolites overlain by a black phosphatic crust. Phosphatic and Proterozoic volcanic clasts within and on this crust form the base of depositional sequence 4B (Jigging Cove Member). Fletcher (2006, p. 48)

reported only 71 m of red mudstone and nodular carbonates at Jigging Cove as his “Redlands Cove Member” and reported higher strata as supposedly lacking nodules and referable to his “Jigging Cove Member.” Fletcher’s (2006) proposed members of the Brigus and Chamberlain’s Brook formation were abandoned by Westrop and Landing (2012; also Landing *et al.*, 2013b) in preference to the members earlier formalized by Landing and Westrop (1998a, b).

The section north of the massive limestone is the type section of the Jigging Cove Member of the Brigus Formation (Landing and Westrop, 1998a, b). The lowest beds are 14 m of pyritiferous, laminated, green mudstone that mark transgression and initial deposition of a dysoxic facies (compare Myrow and Landing, 1992; Landing and Westrop, 2004). Higher beds of the Jigging Cove Member include reddish mudstones with abundant calcareous nodules (*contra* Fletcher, 2006, p. 54 figure 14, where nodules are said to be lacking). The St. Mary’s and Jigging Cove members are restricted to the St. Mary’s-east Trinity axis (Figures 2, 3, 5).

Stop 9. Sunnyside: Avalonian Lower Cambrian Depositional Sequences 2–3 (Cambrian Stage 2) (June 28; EL, SRW)

Location

From the Trans-Canada, take the intersection with the secondary road to Sunnyside village. Drive through the village and park at the east end of high, south-facing road cuts. Descend to the beach on the north shore of Bull Arm fjord and walk ca. 200 m west to the contact of the Random Formation with the overlying basal Bonavista Group (Figure 2, locality Su).

Generalized Stratigraphy and Significance

Stop 9 is important in SE Newfoundland Cambrian stratigraphy as it features a well exposed section from the top of the Random Formation through the upper member (member 4) of the Cuslett Formation (Landing and Benus, 1988a). Sunnyside lies almost on the Placentia–Bonavista axis (Figures 2, 3, 5) and has a very thick Bonavista Group section.

Unconformity and epeirogenic activity are shown at the Random–Petley formation contact (the latter unit is only 3.5 m-thick). There is an absence of the thick sandstones in the upper Random at Hickman’s Harbour and Smith Sound (Hutchinson, 1962) as a result of erosion. The top of the Random features 75 cm of weathered (reddish and calcified) green mudstone. This mudstone has deep, subvertical to horizontal, quartz sand- and carbonate-filled fissures produced by post-depositional and post-cementation fracturing of the Random by regional extension. This epeirogenic extension is interpreted to have produced the St. Mary’s–east Trinity axis and the Avalonian Depositional Sequence 2–3 unconformity (Landing and Benus, 1988b).

Successive tidal-channel fills (SW-flow) with calcareous sandstone and Fe-Mn-impregnated stromatolites compose most of the Petley. The top bed of the Petley is a high-stand carbonate with an *Aldanella*-orthotheidid hyolith fauna.

Stop 10 features the type section of the West Centre Cove Formation (35.4 m-thick)—a reddish, generally structureless (burrow-churned) siliciclastic mudstone-dominated unit with upper gray-green mudstones, and two highstand limestone beds (34.0–35.4 m) separated by 20 cm siliciclastic mudstone (Figure 44). Peritidal conditions (mudcracked planar stromatolites) are shown in the upper limestone bed and feature an abundant *Aldanella*-orthotheidid hyolith fauna of the *Watsonella crosbyi* Zone.

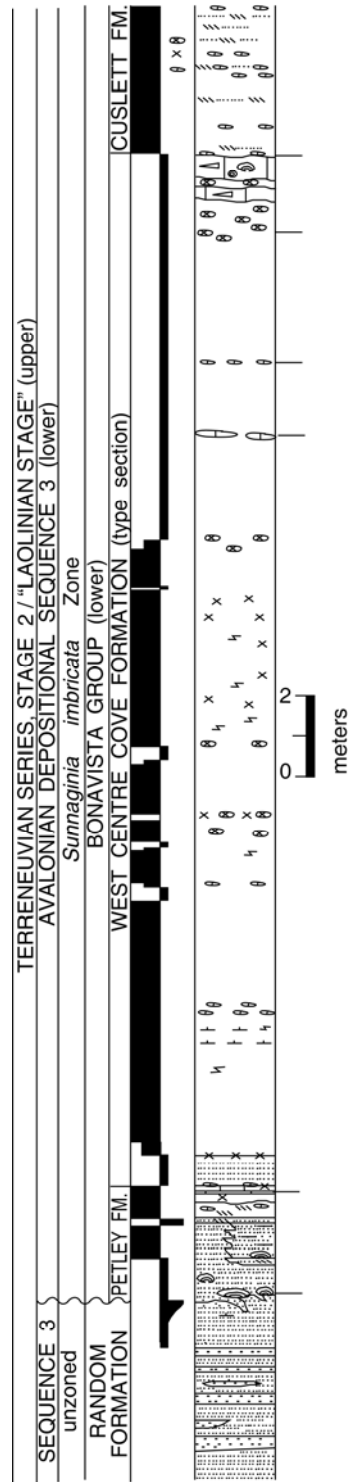


Figure 44. Lower part of Sunnyside succession (Stop 8). Section includes type section of the West Centre Cove Formation. Lithologic symbols explained in Figure 47 caption. Modified from Landing et al. (1988, fig. 30). Short horizontal lines to right of stratigraphic column are microfossil sample horizons.

The presence of *Watsonella crosbyi* Zone faunas in the underlying Chapel Island Formation (Stop 6A) and its persistence through the West Centre Cove Formation and into the lower Cuslett Formation (member 1) at this stop (Stop 10) (Landing and Benus, 1988a) is an indication of the very long time interval represented by this early biomineralized faunal assemblage. Indeed, *W. crosbyi* Zone faunas likely bracketed about 10 m.y. of geological time (Landing and Kouchinsky, 2015; Landing *et al.*, 2015b; Landing, 2015).

The overlying Cuslett Formation can be measured to 235 m above the Random to the hill crest above the road. The first amalgamated nodular limestone that defines the base of member 2 of the Cuslett also lies at the base of the *Camenella baltica* Zone (with the lowest occurrence of *C. baltica*, very abundant *Coleoloides typicalis*, and the lowest hyolithid hyoliths). The base of the *C. baltica* Zone, only marked by the lowest occurrence of *C. baltica* and the persistence of all *Watsonella crosbyi* Zone taxa, is really almost a “non-event” that reflects a linked litho- and biofacies change (Landing and Benus, 1988b; Landing *et al.*, 1989; Landing, 1992). It is certainly not an unconformity (*vide* Steiner *et al.*, 2007).

Stop 10. Smith Point: Trans-Avalonian Depositional Sequence 3–4 Unconformity and Earth’s Longest “Worm Reef” (June 28; EL, SRW)

Location

Turn east onto Route 230 from the Tran-Canada and drive through Clarendville, Milton, and Georges Bank villages. At the intersection with Rte. 232, turn east on Rte. 232 through Harcourt village and into Fosters Point village (discussed below). Turn north (right) onto the steep(!) road to the government wharf (Figure 2, locality SP). The condensed trilobitic limestone

at the base of the Brigus Formation in this gently west-dipping section lies under the old wooden building just east of the wharf. The type section of the Fosters Point Formation is exposed immediately to the east, and the Brigus Formation is exposed under the wooden building and crops out to the west along the shore.

Generalized Stratigraphy and Significance

The upper part of member 4 (green mudstone) of the Cuslett Formation (Landing and Benus, 1988b) underlies the interbedded nodular red limestone and red siliciclastic mudstone of the lower Fosters Point Formation (7 m-thick) (Figure 45). The Fosters Point is the highstand carbonate of the Bonavista Group. This is the type section of the Fosters Point Formation (Landing and Benus, 1988b)—“Fosters Point” is the traditional name for the settlement on the road immediately above the Smith Point section (Jenness, 1963, Map 1130A).

The abundance and density of nodular carbonate increases upward in the Fosters Point Formation to form a griotte. Shell hash (primarily wave-sorted *Coleoloides* conchs with hyolith fragments) infills the depressions, just as conchs occur in depressions in the much younger Mesozoic Ammonitico Rosso (e.g., Jenkyns, 1974). These distributions of shell hash indicate that the “nodules” had depositional relief and were likely microbial mud mounds (these non-laminated structures are called “stromatolites” in Fletcher (2006, pl. 25). Wave sorting and orientation of the Fosters Point conchs suggest shallow water deposition not only for the Fosters Point, but also for deposits such as the “Ammonitico Rosso” which are traditionally regarded as “deep-water” (e.g., Jenkyns, 1974).

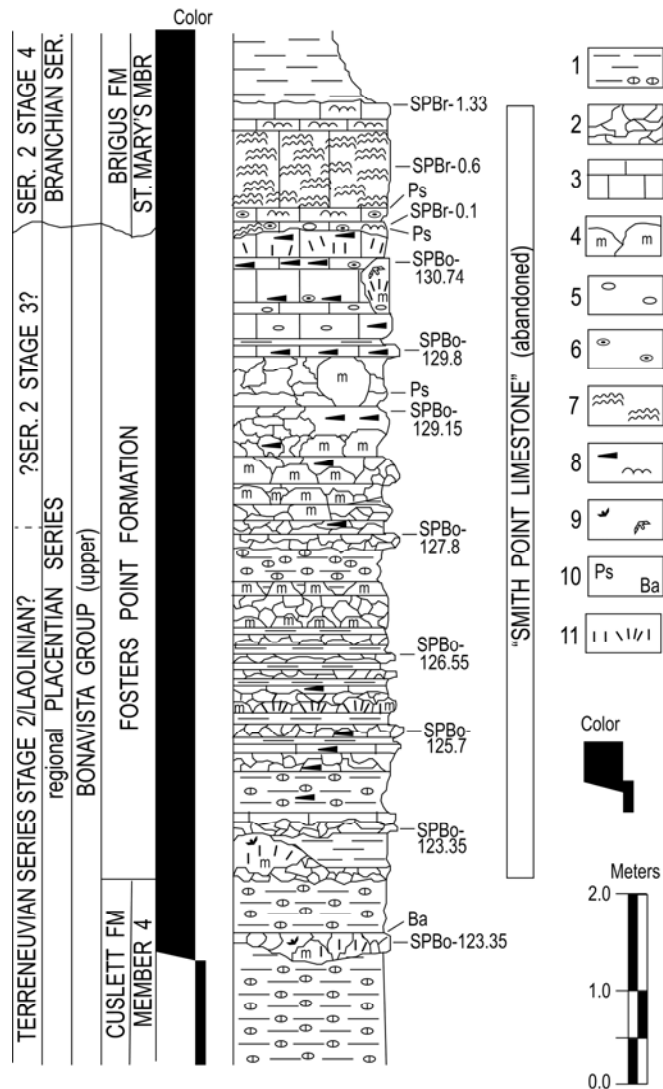


Figure 45. Detailed lithology of type section of *Fosters Point Formation* (Stop 9). Modified from Landing et al. (1988, fig. 31). Key: 1, siliciclastic mudstone, structureless (bioturbated) with calcareous nodules; 2, nodular limestone, likely becoming structureless microbial mounds in upper half of formation; 3, bedded limestone with stylonitic bedding; 4, calcareous microbial mud mounds; 5, intraclasts in limestone beds; 6, oncoids, frequently Fe-Mn-impregnated; 7, planar to *Collenia*-type stromatolites, frequently Fe-Mn-impregnated; 8, hyoliths (upper), trilobites (lower); 9, birdseye structures (left) and stromataxis (right) in microbial mud mounds; 10, bluish-weathering pseudospar (Ps) and barite crystals (Ba); 11, in situ vertical conchs of *Coleoloides*. Sample numbers (e.g., SPBo-123.35) are meters above top of Random Formation or (e.g., SPBr-1.33) meters above base of Brigus Formation and above unconformable contact with top of Bonavista Group (*Fosters Point Formation*).

Larger mud-mounds at 130 m contain *in situ* *Coleoloides* tubes that radiate to the margins of nodules. These nodules commonly have geopetal calcite spar- and calcareous mudstone-filled stromataxis structures.

The top of the Fosters Point is a lenticular bed with *in situ* *Coleoloides* that persists across SE Newfoundland to Cape Breton Island and into eastern Massachusetts. It seems to comprise the most geographically widespread “worm” reef known. Comparable facies occur in the coeval, thin (2.0 m) Home Farm Member of the Hartshill Formation in Avalonian England (Brasier *et al.*, 1978; Brasier and Hewitt, 1979).

The top bed of the Fosters Point Formation at Stop 10 shows ca. 15 cm of erosional relief, and the oncolitic trilobite wackestone of the basal bed (1.5 m-thick) of the Brigus Formation wedges out on this surface. Locally limonitized planar and “*Collenia*”-type stromatolites dominate parts of the lower bed of the Brigus Formation. These true stromatolites appear only in the basal Brigus onlap bed at Stop 10, a unit assigned by Fletcher (2006, pl. 24) to the top of his “Smith Point Formation”—a designation abandoned by Landing (1996a; also Landing and Westrop, 1998a, b). The oldest Avalonian trilobites of the *Callavia broeggeri* Zone (upper Stage 3) appear at this Avalonian Depositional Sequence 3–4A unconformity, but the well cemented nature of the rocks has long prevented comprehensive documentation of fossils from this interval (Hutchinson, 1962; Fletcher, 2006).

Red, purplish-red, and green siliciclastic mudstones higher in the lower Brigus Formation (St. Mary’s Member) crop out as low shore cliffs west of the wooden building. A greenish, fining-upward, glauconitic sandstone (23.25–23.85 m) with limestone and phosphatic pebbles at its base and an abrupt change from the *Callavia broeggeri* Zone into a fauna with

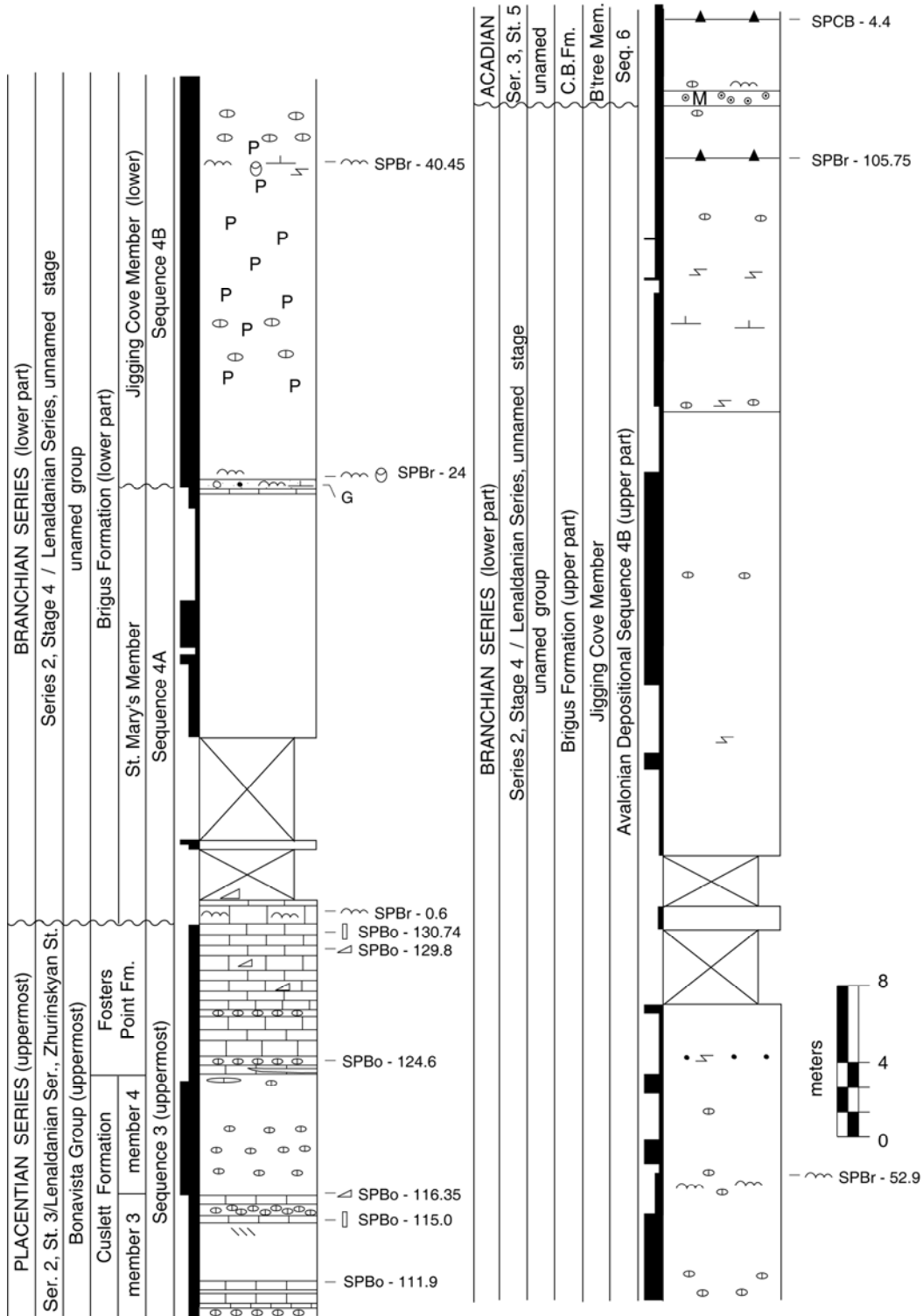


Figure 46. Upper Series 2, Stage 4 (upper Lenaldanian Series)–lower Series 3, Stage 5 at Smith Point, western Trinity Bay (Stop 9). Abbreviations B'tree M., Braintree Member; C.B. Fm.,

Chamberlain's Brook Formation. Explanation of symbols in Figure 8. Modified from Landing and Westrop (1998b, fig. 16.).

Myopsostrenua marks the base of the upper Brigus Formation (Jigging Cove Member) and the Avalonian 4A–4B depositional sequence boundary (Landing and Westrop, 1998b; Figures 5, 46).

Stop 11A. Keels: Regional Cambrian Stage 2 Biostrome and Avalonian Cool-Water Analogues of Tropical Archaeocyaths (June 29; EL)

Location

The thickest known Bonavista Group section in SE Newfoundland is in a north-plunging syncline at Keels at the tip of the Bonavista Peninsula (Figure 2, locality K). The locality is reached by driving down Route 235 from the intersection of Routes 230 and 235 at Southern Bay village to Kings Cove village. Follow Route 235 to Broad Cove and Keels village. Park at the west side of Keels village at the head of the cove south of the government wharf. Walk to low, east-dipping outcrops at the SE end of the cove for Stop 11A (Figure 47).

Generalized Stratigraphy and Significance

The Keels syncline is on the inner platform, and the Avalonian oldest cover units (Rencontre and Chapel Island formations) are absent (Figure 48). The Keels section at Stop 11A is measured from the top of the Random Formation through the Petley Formation (38 m) and West Centre Cove Formation (79 m). Overlying strata higher in the Cuslett Formation comprise Stop 11B.

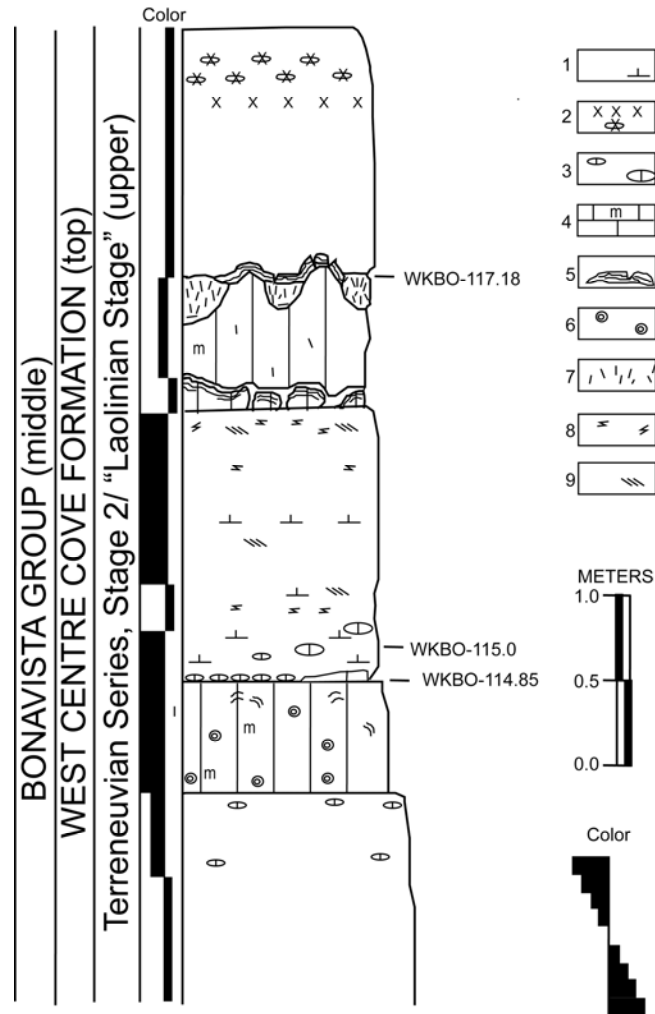


Figure 47. *Stratigraphy of “Ladatheca” cylindrica-microbialite buildups at top of West Centre Cove Formation, middle Bonavista Group, on west limb of Keels syncline (Stop 11A). Modified from Landing et al. (1988, fig. 32)*

The low outcrop at the SE end of the cove includes two massive beds of limestone separated by a green mudstone at the top of the West Centre Cove Formation—an interval 113.0–117.18 m above the Random Formation (Figure 47). Keels, Summerside, and Clifton (Figure 2, localities K, Su, Cl) are localities in the Placentia–Bonavista axis that have two, rather than one, highstand carbonate bed at the top of the West Centre Cove Formation.

The depositional relief on the upper bed has been increased by movement on slip cleavage. It consists of swells and depressions of a pinkish-green microspar. Lensing, limonite- and silica-impregnated, planar and digitate microbial mats infill depressions and have vertically embedded, radiating “*Ladatheca*” *cylindrica* conchs. A few of these conchs are also vertically

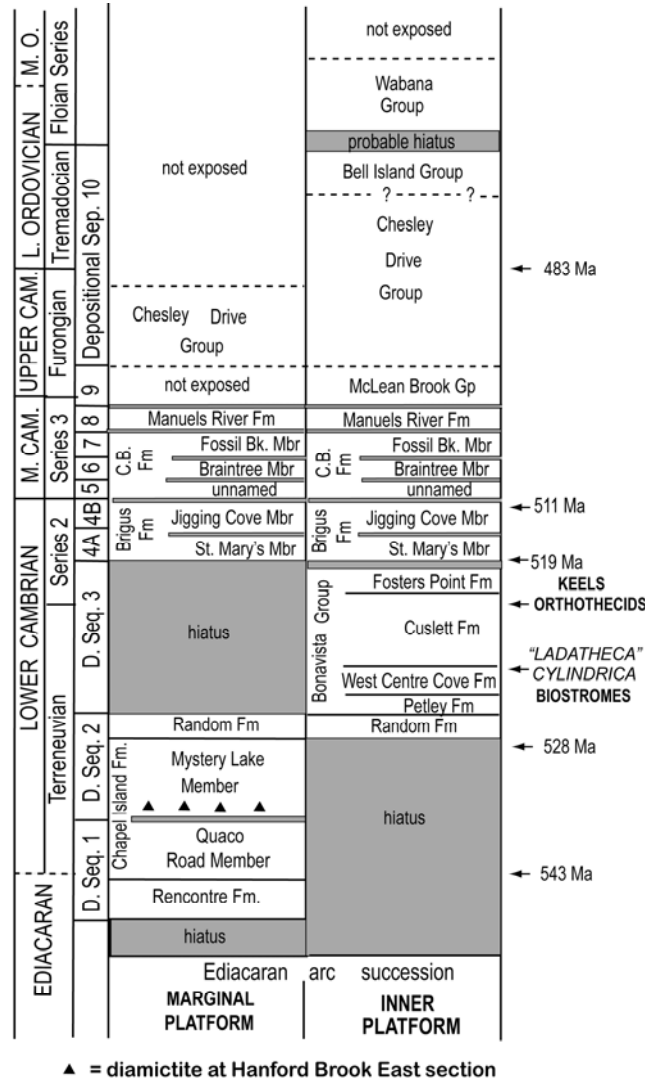


Figure 48. Generalized stratigraphy of Avalonian inner platform along the Placentia-Bonavista axis shows location of “*Ladatheca*” *cylindrica*-microbialite buildups at top of West Centre Cove Formation (Stop 11A) and in situ “*Allatheca*” *degeeri* s.l. conchs in the upper Cuslett Formation (Stop 11B) in the Keels syncline. Modified from Landing and Kröger (2012, fig. 2).

embedded in the microspar. These vertical occurrences of “*L.*” *cylindrical* in peritidal mud mounds (this stop) and in deeper-water siliciclastic mudstone (Stop 6A) indicate the life position of this eurytopic species (Landing, 1994; Landing and Westrop, 2004). Archaeocyaths are common in West Gondwana and on other tropical paleocontinents, while *in situ* “*Ladatheca*” and *Coleoloides* in mud mounds are their ecological homologues in the shallow shelf successions of Avalonian Newfoundland (Stops 10, 11A), Cape Breton Island (EL., unpub. data) and England (Brasier and Hewitt, 1979).

Stop 11B. Keels: *In Situ* Stage 3 Hyoliths (June 29; EL, BK)

Location

The locality forms the very northern tip of the Keels syncline. It is reached by driving back through Keels, turning left into the gravel pit, and parking at the north end. A shore path extends to the north end of the exposure.

Generalized Stratigraphy and Significance

There is almost 100% exposure of the Cuslett Formation along the east shore. Red, purple-red, purple, and green siliciclastic mudstone alternations through the lower section represent more proximal to distal facies, respectively (Landing and Benus, 1988a, b; Landing *et al.*, 1989, 2013a; Myrow and Landing, 1992). Two distinct intervals of dark gray mudstone (267.75–248.5 m and 262.2–264.6 m) appear in green mudstone with abundant small traces and chlorite grains (after glauconite) near the top of the section.

The lower gray mudstone has nodules (samples KBo-243.0, -243.75) with a *Camenella baltica* Zone fauna (*Rhombocorniculum cancellatum*, *Halkieria fordi*). This is another possibly Atdabanian-equivalent (or older) fauna without trilobites that further demonstrates that a “FAD of trilobites” is inappropriate as a base for Cambrian Series 2, Stage 3 (*vide* Peng and Babcock, 2005, 2011).

The upper mudstone has pyritized, apically septate conchs of the orthothecid hyolith “*Allatheca*” *degeeri* Holm (Figure 49). Landing and Kröger (2012) noted that 46 of the observed conchs, including the largest (10 cm-long), are horizontally embedded, and show a bimodal (primarily NE and SW) oscillatory wave-determined distribution, a feature that accords with relatively shallow-water deposition of the organic-rich mudstone.

A few smaller specimens (7 conchs, to 5 cm-long) are oriented aperture-down at angles of 45°–90° to the horizontal. Simple arithmetic calculations based on conch wall and sea-water density, conch thickness, and the potential role of a hypothetical gas located in the camera behind the apical septum rule against the possibility of the “*A.*” *degeeri* individuals as being neutrally buoyant or free-swimming and living like a shelled cephalopod (Landing and Kröger, 2012). Slabbing of dark mudstones provided no evidence for the occurrence of the aperture-down conchs in muddy tempestites or on submarine firmgrounds to which the animals may have been attached. Thus, the vertically embedded conchs suggest a habitat below the sediment-water interface and a mode of life habit as a detritivore somewhat similar to a scaphopod (Landing and Kröger, 2012). Although hyoliths are commonly figured as reclined, epibenthic, filter-feeding, “shell-lugging” animals, a broader range of modes of life is more likely. This is suggested by the scaphopod-like, infaunal life habit of the Keels orthothecids. In addition, the scars on the operculum of the hyolith *Cupitheca* resemble those of juvenile brachiopods, rugose corals, and

gryphaeids, and indicate an attached, epibenthic life mode, at least for juveniles (Skovsted *et al.*, 2016).

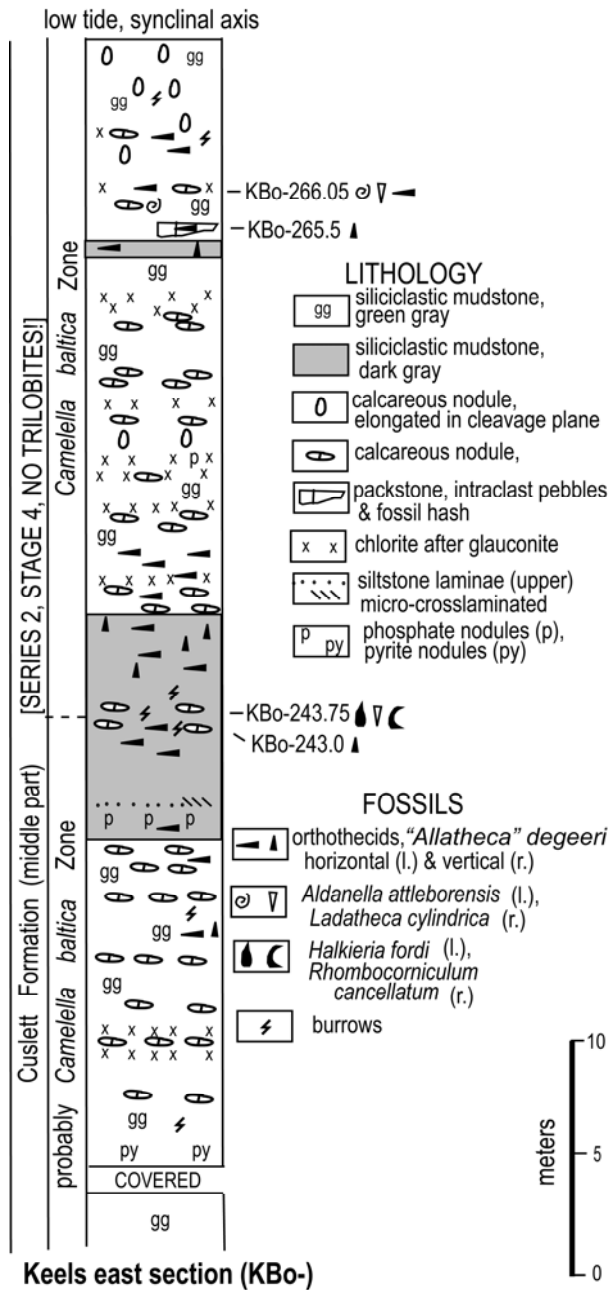


Figure 49. Stratigraphy of in situ "*Allathea*" *degeeri* s.l. conchs in the upper Cuslett Formation, upper Bonavista Group; specimens occur in highest strata of the Keels syncline (Stop 11B). Modified from Landing and Kröger (2012, fig. 3).

ACKNOWLEDGMENTS

This field trip guide is dedicated to the memory of two Memorial University of Newfoundland faculty—to Michael M. Anderson who genially dispersed his knowledge of SE Newfoundland geology and to Hank Williams, dean of synthesizers of Appalachian Mountain geology.

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