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First evidence for Cambrian glaciation provided by sections in Avalonian New Brunswick and Ireland: Additional data for Avalon–Gondwana separation by the earliest Palaeozoic

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ABSTRACT

The first evidence for Cambrian glaciation is provided by two successions on the Avalon microcontinent. The middle lowest Cambrian (middle Terreneuvian Series and Fortunian Stage–Stage 2 boundary interval) has an incised sequence boundary overlain by a fluvial lowstand facies and higher, olive green, marine mudstone on Hanford Brook, southern New Brunswick. This succession in the lower Mystery Lake Member of the Chapel Island Formation may be related to melting of an ice sheet in Avalon. The evidence for this interpretation is a muddy diamictite with outsized (up to 10 cm in diameter), Proterozoic marble and basalt clasts that penetrated overlying laminae in the marine mudstone. That eustatic rise was associated with the mudstone deposition is suggested by an approximately coeval rise that deposited sediments with *Watsonella crosbyi* Zone fossils 650 km away in Avalonian eastern Newfoundland. A sea-level rise within the *Watsonella crosbyi* Chron, at ca. 535 Ma, may correspond to a unnamed negative ^{13}C excursion younger than the basal Cambrian excursion (BACE) and the ZHUCE excursion in Stage 2 of the upper Terreneuvian Series. Cambrian dropstones are now also recognized on the northern (Gander) margin of Avalon in continental slope–rise sedimentary rocks in southeast Ireland. Although their age (Early–Middle Cambrian) is poorly constrained, dropstones in the Booley Bay Formation provide additional evidence for Cambrian glaciation on the Avalon microcontinent. Besides providing the first evidence of Cambrian glaciation, these dropstone deposits emphasize that Avalon was not part of or even latitudinally close to the terminal Ediacaran–Cambrian, tropical carbonate platform successions of West Gondwana.

1. Introduction

Reliable evidence for Cambrian glaciation has not been documented, and the period has been regarded as a time in which continental glaciation is unknown (e.g. Hambrey and Harland, 1981, p. 970). Various authors have proposed that Cambrian and Early Ordovician eustatic rises and falls were caused by the waning and waxing of continental glaciers (e.g. Erdtmann and Miller, 1981; Fortey, 1984), but have provided no physical evidence for these assumed glaciations. As noted by Moczydłowska-Vidal and Vidal (1997), earlier beliefs that late Proterozoic glaciation(s) may have continued into the Cambrian (e.g. Harland, 1964) were shown to be incorrect as these glaciations were shown to be pre-Ediacaran with the use of improved biostratigraphic and geochronologic correlation techniques for the Precambrian–Cambrian boundary interval.

Bertrand-Safarti et al. (1995, 1997) and Trompette (1996) argued that glaciation persisted from the Proterozoic into the Early Cambrian in West Africa. However, Trompette (1996) provided no justification for correlating a Cryogenian cap carbonate in the Pan-African orogen of central Algeria with a distant, thin, undeformed carbonate succession with Early Cambrian small shelly fossils in Senegal (see Culver et al., 1996). Similarly, the questionably Ediacaran remains described by Bertrand-Safarti et al. (1995) succeed Cryogenian glacials and are overlain by carbonate with Early Cambrian fossils (Culver et al., 1996), with no evidence for stratigraphic continuity. Bertrand-Safarti et al. (1995) assigned a Cambrian age to the Fersiga diamictite, in part due to the presence of the fossils. However, subsequent work has indicated that the fossils are pre-Ediacaran and middle Cryogenian in age (MacGabhann, 2007a) and that the Fersiga diamictite actually correlates with the Elatina diamictite in South Australia (Shields et al., 2007a,b), at the top of which is the global stratotype for the Ediacaran Period.

In any case, the south Moroccan margin of West Gondwana shows that the Pan-African orogen is unconformably overlain by characteristic tropical carbonate and evaporate deposits of terminal Ediacaran(?) to middle Early Cambrian age (e.g. Geyer and Landing, 1995, 2006; Michard et al., 2008). Thus, West Gondwana is an unlikely region to have undergone glaciation through this time interval.

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A far more likely area of Ediacaran glaciation was in an arcuate belt that extends from central and southern Kazakhstan through Kyrgyzstan and into northwest China. Chumakov's (2009) useful review notes that "Baykonurian glaciation" diamictites overlie or are interbedded with bedded sedimentary rocks with such late Ediacaran forms as *Vendotaenia* and *Sabellidites*. However, incomplete information is available on the upper contacts of these Ediacaran glaciogenic rocks, and Korolev and Maksumova (1984) distinguish an unconformity between the Baykonurian glacials and overlying Lower Cambrian sedimentary rocks. These Cambrian rocks lack glacial deposits and frequently comprise tropical carbonate platform facies a short stratigraphic distance above the glacials.

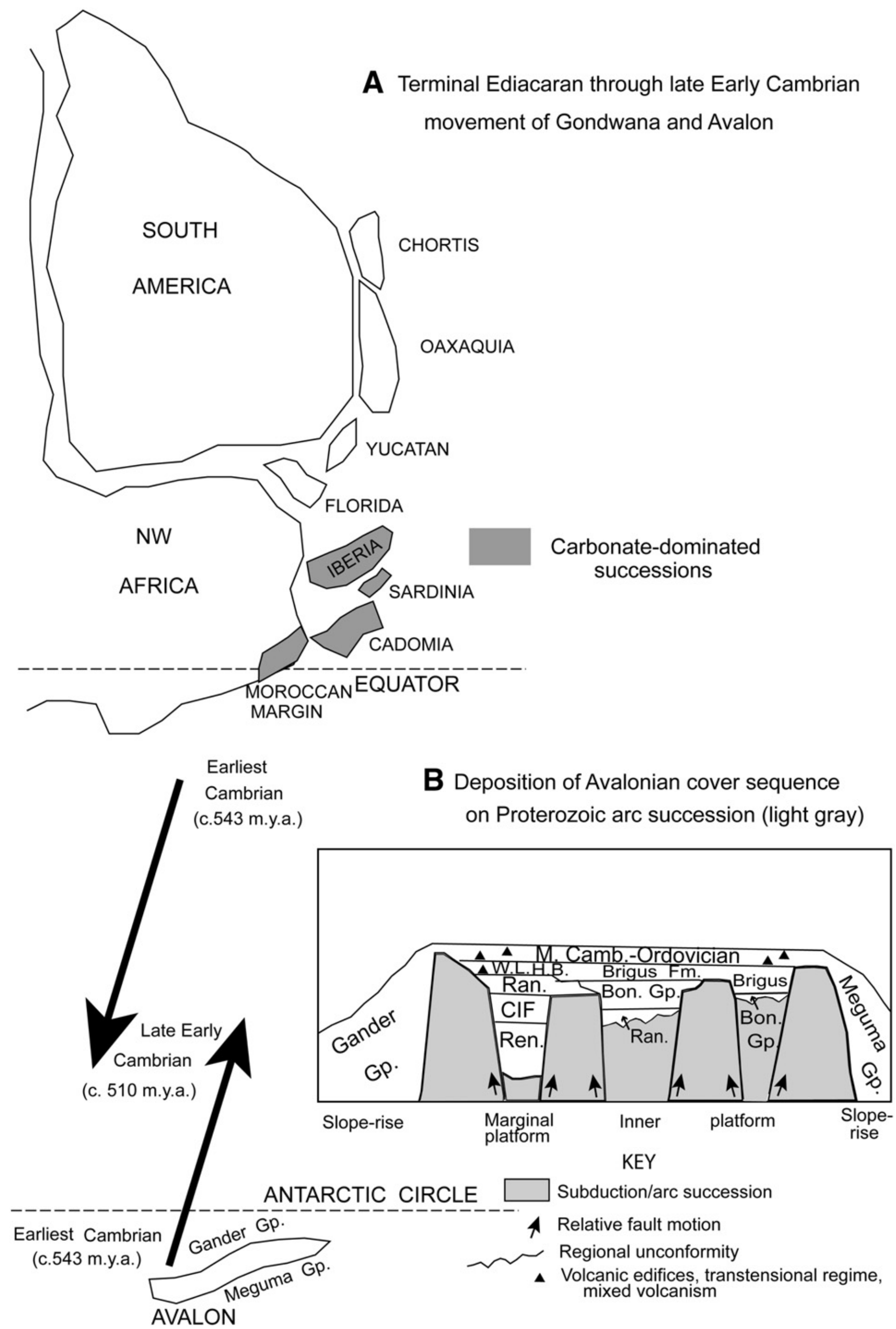

Fig. 1. Terminal Ediacaran–late Early Cambrian (terminal Series 2) palaeogeography and depositional tectonics of Avalon. (A) Early Cambrian palaeogeography of Avalon, Gondwana, and Peri-Gondwanan terranes (modified from Keppie et al., 2003, fig. 2) with northerly transport of the Avalon microcontinent and southerly transport of Gondwana into comparable mid-latitudes by the terminal Early Cambrian. (B) NW–SE transect (in modern coordinates) across the Early Paleozoic Avalon microcontinent, figure shows successive southeast movement of fault-defined depocenters through the Cambrian. Abbreviations: Bon. Gp., Bonavista Group; CIF, Chapel Island Formation; Gp, Group; H.B., Hanford Brook Formation; Ran., Random Formation; Ren., Rencontre Formation; W.L., Wade's Lane Formation. Volcanic edifices on marginal platform include late Early Cambrian Wade's Lane Formation at Beaver Harbour (Fig. 2, loc. BHr, see Landing et al., 2008), Middle Cambrian basalts in the Malignant Cove block, mainland Nova Scotia (Fig. 2); late Early Cambrian and Middle Cambrian Escasoni and Gregwa Formations in Cape Breton Island (Hutchinson, 1952), and middle Middle Cambrian basalts in the Manuels River Formation, Red Bay, Burin Peninsula, eastern Newfoundland (E. Landing, unpublished data). Inner platform volcanic edifices are in southeastern Newfoundland and include a Middle Cambrian edifice, probably in the Manuels River Formation at Cape Dog, St. Mary's Bay (Hutchinson, 1962, Section 11), and in the upper Manuels River Formation at McLeod Point and Little Ridge, Trinity Bay (Hutchinson, 1962, Sections 8, 9).

Cocks and Torsvik's (2002, fig. 3) palaeogeographic map of the Middle and Late Cambrian places the north-central African part of Gondwana on the South Pole and locates a number of "peri-Gondwanan" regions (i.e., Iberia and Sardinia) south of the Antarctic Circle. However, their reconstruction is inconsistent with the facts that terminal Ordovician, not Middle–Late Cambrian, glacial deposits are known in West Africa (e.g. Le Heron, 2007), while thick, tropical, carbonate platform deposits that range from late Early Cambrian age in southern Spain (Ossa-Morena and Galician zones/terranes) to middle Middle Cambrian (Cantabrian Zone/terrane) in northern Spain blanketed Iberia in the Cambrian (Liñán et al., 2004; Palacios et al., 2008). Similarly, the middle–late Early Cambrian of southwest Sardinia featured development of a thick carbonate platform succession (Pillola, 1991).

Evidence for Cambrian glaciation has now been obtained in southern New Brunswick and southeastern Ireland in the course of a regional study of rock successions on the Avalon microcontinent (e.g. Landing, 1994a, 1996a, 2004; Landing and Westrop, 1998). The Avalon microcontinent is now a fault-segmented terrane in the Acadian–Caledonian orogen of the northern Appalachian Mountains, southern Britain, and central continental Europe (see original proposal of the 'Avalon Platform' by Rast et al., 1976). The newly discovered glacial sedimentary rock can be biostratigraphically bracketed in New Brunswick to an interval within the earliest Cambrian. The relative sea-level (i.e., palaeobathymetric) trends recorded in the lithofacies that bracket this glaciogenic deposit suggest regression and transgression, and, perhaps, that a eustatic fall–rise couplet occurred in this time interval as a result of glaciation and deglaciation. Finally, the glacial deposits on the Avalon platform in southern New Brunswick and in continental slope–rise deposits on the northern (Wexford terrane) margin of Avalon in Ireland provide further evidence that Avalon was latitudinally distant and climatically distinct from tropical West Africa by the end of the Ediacaran (Landing, 1994b, 1996a, 1998, 2003, 2005). Accurate palaeogeographic maps must reflect this distribution of palaeocontinents (Fig. 1), rather than continue to repeat variations of the Scotese and McKerrow (1990) and McKerrow et al. (1992) palaeogeographic maps that show Avalon as a part of or close to West Gondwana through the Cambrian, with a separation only in the Ordovician (e.g. Scotese and McKerrow, 1990; McKerrow et al., 1992; Fortey and Cocks, 1992, 2003; Álvaro et al., 2003; Linnemann et al., 2008; Palacios et al., 2008).

2. Geological setting

2.1. Avalonian cover succession

Terminal Ediacaran(?)–Ordovician sedimentary rocks comprise a weakly metamorphosed (low chlorite grade) cover sequence that unconformably overlies a late Precambrian subduction/arc succession and older rocks in coastal, southern New Brunswick and other regions on the platform of the Avalon microcontinent (e.g. Rast et al., 1976; Landing and Westrop, 1998; Keppie et al., 2003). Most cover successions in southern New Brunswick span the earliest Cambrian, and were deposited in syndepositional, fault-bounded depocenters on the Avalonian marginal platform (Figs. 1B and 2). Marginal platform cover successions in northern Wales, the Burin Peninsula of eastern Newfoundland, Cape Breton Island, and southern New Brunswick are similar in that they are thick successions (ca. 3 km, locally) that include lower rift facies (subaerial conglomerates to marginal marine facies); and overlying siliciclastic, wave-dominated shelf facies followed by a tidalite quartz sandstone. In Avalonian North America, a unified nomenclature has been applied to these three intervals: respectively, the Rencontre, Chapel Island, and Random Formations (e.g. Landing and Westrop, 1995; Landing, 1996a, 2004). Inner platform successions, which have the Random Formation or younger Cambrian rocks as the lowest cover unit, are poorly represented in Avalonian New Brunswick, and are presently known only from the eastern locality at Cradle Brook (Landing, 1996b) (Figs. 2 and 3).

Controversy surrounds the present location of the deep-water, continental slope–rise packages that would have existed marginal to the Avalon microcontinent. However, Keppie et al. (2003) provide a persuasive argument that the deep-water siliciclastic-dominated, Ediacaran(?)–Lower Palaeozoic Gander and Meguma Groups, which lie northwest and southeast, respectively, of the Avalon platform in North America, are Avalonian slope–rise packages. The Gander and Meguma Groups provided the basis for the Gander and Meguma Zones of the central and southern Appalachian Mountains, respectively, in North America (e.g. Williams, 1978). Further east, terminal Ediacaran (?) and Cambrian–Ordovician siliciclastic successions of southwestern Ireland (Wexford terrane) and the Brabant and other anticlinoria of central Europe are regarded as continental slope–rise successions

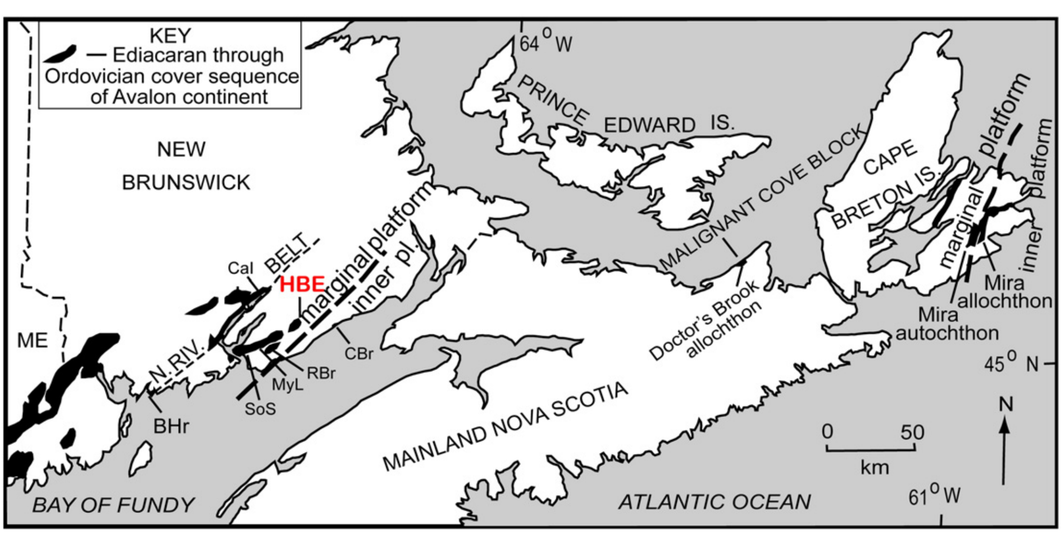

Fig. 2. Marginal and inner platforms in Avalonian Maritime Canada. Black areas are synclinal and down-faulted outliers of terminal Ediacaran–Ordovician cover sequence. Abbreviations: BHr, Beaver Harbour; CBr, Cradle Brook; SoS, HBE, Hanford Brook East section (Fig. 4); IS., Island; ME, Maine; MyL, Mystery Lake; N. Riv. Belt., New River Belt; RBr, Ratcliffe Brook section; SoS, Somerset Street. Note: The Mira River separates the Mira Autochthon (on west side of river) from the thrust-transported Mira Allochthon (to east and south of Mira River).

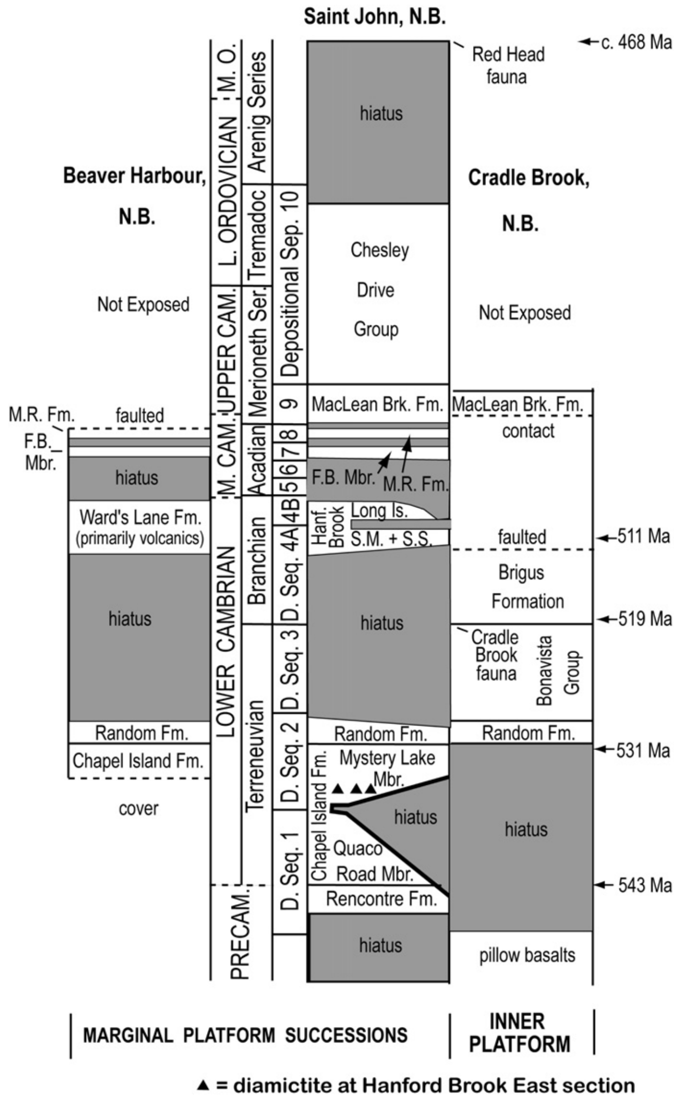

Fig. 3. Marginal platform successions in Beaver Harbour and Saint John, New Brunswick, areas, compared with inner platform succession at Cradle Brook. Stratigraphic column scaled proportional to Cambrian–Ordovician geochronology (see Tucker and McKerrow, 1995; Landing et al., 1998, 2003); grey = hiatus corresponding to regional sequence boundaries on Avalonian marginal and inner platforms. Abbreviations: F.B. Mbr., Fossil Brook Member of Chamberlain's Brook Formation; Fm., Formation; Hanf. Brook, Hanford Brook Formation; Long Is., Long Island Member; M.R. Fm., Manuels River Formation; S.M. and S.S., St. Martin's and Somerset Street Members respectively. Cradle Brook and Red Head faunas in Landing (1996a) and Landing et al. (2003).

marginal to northwest and southeast Avalon, respectively, and occupy palaeogeographic positions comparable to the Gander and Meguma Groups (e.g. Geyer et al., 2008; Linnemann et al., 2008; Fig. 1).

2.2. Southern New Brunswick sections in the Chapel Island Formation

Sections that extend from the volcanic-dominated Coldbrook Group (ca. 600 Ma, see Currie, 1987; Barr and White, 1999) through the unconformably overlying Rencontre Formation, both members (Quaco Road and overlying Mystery Lake) of the Chapel Island Formation, and the Random Formation occur east of Saint John New Brunswick. These successions lie along Ratcliffe and Hanford Brooks (Hayes and Howell, 1937; Alcock, 1938; Fig. 2, localities RBr and HBE). Further east in Saint John, the lowest cover unit is the upper member (Mystery Lake Member) of the Chapel Island Formation. In Saint John, the underlying Quaco Road Member and Rencontre Formation are absent at an unconformity with Ediacaran (Coldbrook Group volcanics) and older rock units (Landing, 2004; Fig. 2, localities MyL and SoS).

The long, gently west-dipping Hanford Brook section has been divided into three parts. These include an eastern section, "Hanford Brook East" (HBE), that spans the upper Coldbrook Group through the lowest Random Formation. The "Hanford Brook Central" (HBC) section ranges through the Random to the base of the unconformably overlying Hanford Brook Formation, and the "Hanford Brook West" (HBW) section includes the Hanford Brook Formation through Middle Cambrian (Landing and Westrop, 1998, figs. 22–24; Landing, 2004, figs. 9 and 10).

The middle Chapel Island Formation at Ratcliffe Brook and Hanford Brook East is characterized by wave-dominated, feldspathic sandstone deposits that are sharply overlain by conglomerate beds dominated by Coldbrook Group volcanic and hydrothermal quartz pebbles. These coarse-grained beds are then succeeded by dark olive green mudstones. The base of the conglomerate unit cuts down approximately 1.5 m on Ratcliffe Brook into feldspathic sandstone. This unconformity is regarded as the contact between a depositional sequence boundary between the Quaco Road Member and overlying Mystery Lake Member (Landing, 1996a, 2004). The contact between the two members is covered at the Hanford Brook East section, and lies in a covered interval between the highest sandstone and lowest conglomerate at ca. 159 m (Fig. 4).

3. Sedimentology of the Chapel Island Formation dropstone bed and bracketing units

3.1. Upper Quaco Road Member

3The Quaco Road Member of the lower Chapel Island Formation forms the 32.5–ca. 159 m interval of the Hanford Brook East succession (Landing and Westrop, 1998, fig. 22; Landing, 2004, fig. 9). The Quaco Road Member is a deepening–shoaling succession, with feldspathic, purple-red, hummocky cross-stratified sandstones (115.8–127 m) composing the deepest facies on a marine shelf (Landing and Westrop, 1998; not shown in Fig. 4). These are succeeded by purple to pink, medium- to thick-bedded, bidirectional, trough cross-bedded (troughs to 50 cm), coarse-grained feldspathic sandstones with quartz and feldspar granules deposited by probable tidal currents as foreshore bars (127–136.75 m). The top of the Quaco Road Member (Fig. 4, ca. 158 m) is purple to red, thin- to medium-bedded quartz arenites with minor feldspar, bidirectionally oriented cross-bed sets (to 10 cm thick), and local wave-rippled surfaces.

3.2. Lowest Mystery Lake Member fluvial conglomerate and sandstone

The Mystery Lake Member is a thick interval (ca. 159–335 m) in the Hanford Brook East section (Landing, 2004, fig. 9). The lowest part (Fig. 4, 160–162.75 m) is partly covered, but has two beds of clast-supported, red weathering, purple conglomerate with granules to rare small pebbles of hydrothermal quartz, feldspar, and Coldbrook volcanics in a feldspathic matrix. The lack of evident sedimentary structures makes their mode of deposition indeterminable.

The interval 163.75–173.9 m is largely a trough cross-bedded, purple-red, pebble conglomerate facies with coarse-grained sandstone lenses (Figs. 4 and 5A). The cross-bedding sets are 20–50 cm in thickness, show fining-up trends, and are cross sections of channels. The transport direction was approximately north. This conglomerate and sandstone interval, with large clasts of Coldbrook rhyolite and basalt (to 10 × 5 cm in diameter), is interpreted as a braided stream facies with high gradients and small channels.

3.3. Strata above the fluvial deposits

The strata immediately above the fluvial conglomerates and sandstones record a sharp lithologic break to a dark olive green mudstone-

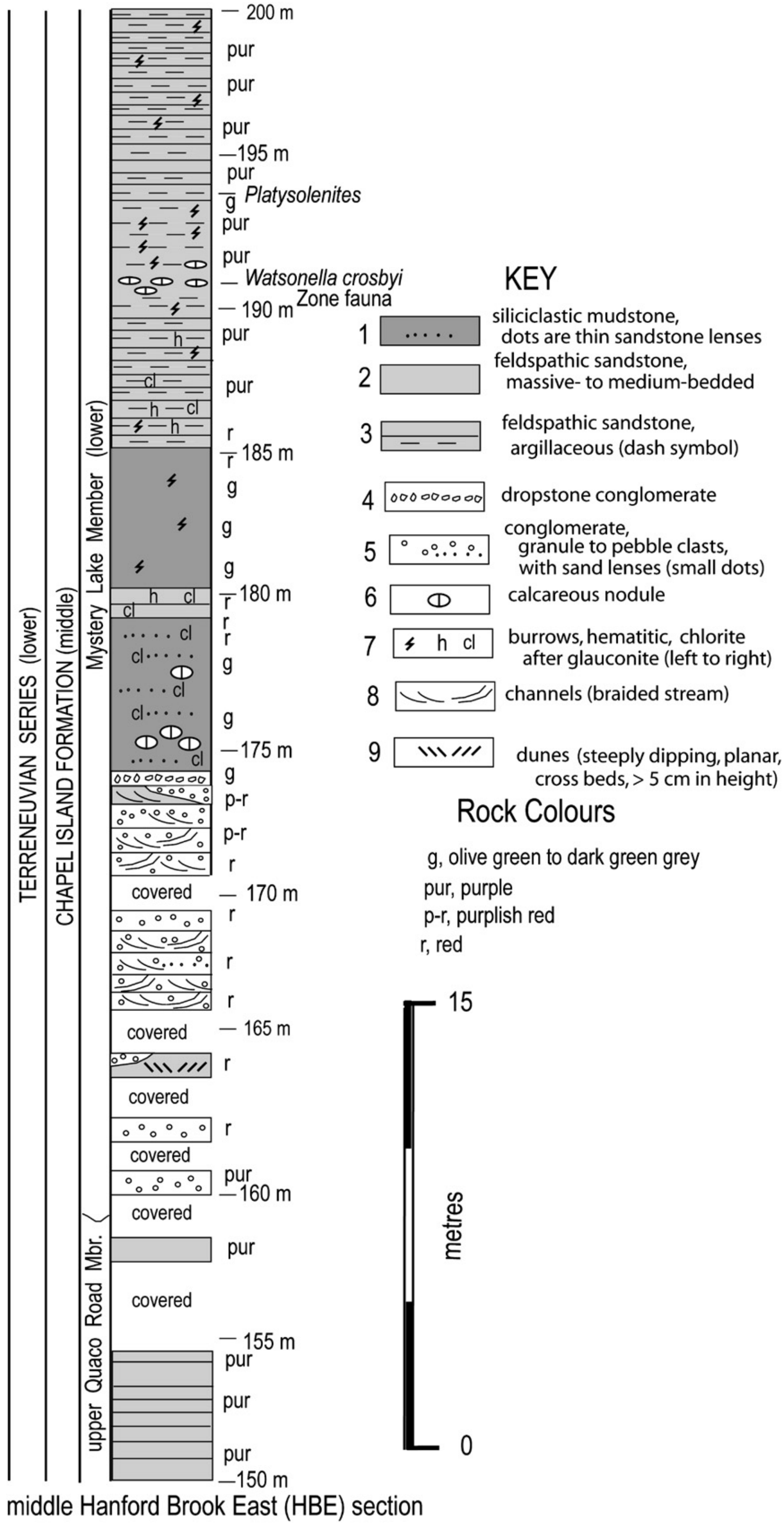

Fig. 4. Stratigraphic succession of the middle (150–200 m) of the Hanford Brook East (HBE) section (378 m-thick) (for complete Hanford Brook East sections see Landing and Westrop, 1998, Fig. 22; Landing, 2004).

dominated interval (173.9–177.75 m) that is thin-bedded and has large calcareous nodules (Figs. 4 and 5B). No trace or body fossils have been recognized in the mudstone or nodules. Numerous centimetre-thick lenses of micaceous quartz arenite occur through the mudstone. Rounded chlorite grains that are ooidal and botryoidal are abundant in the sandstone lenses, and appear to have been metamorphosed from glauconite. If so, the glauconite is an indicator that the mudstone was deposited under marine conditions (e.g. Garzanti et al., 1989).

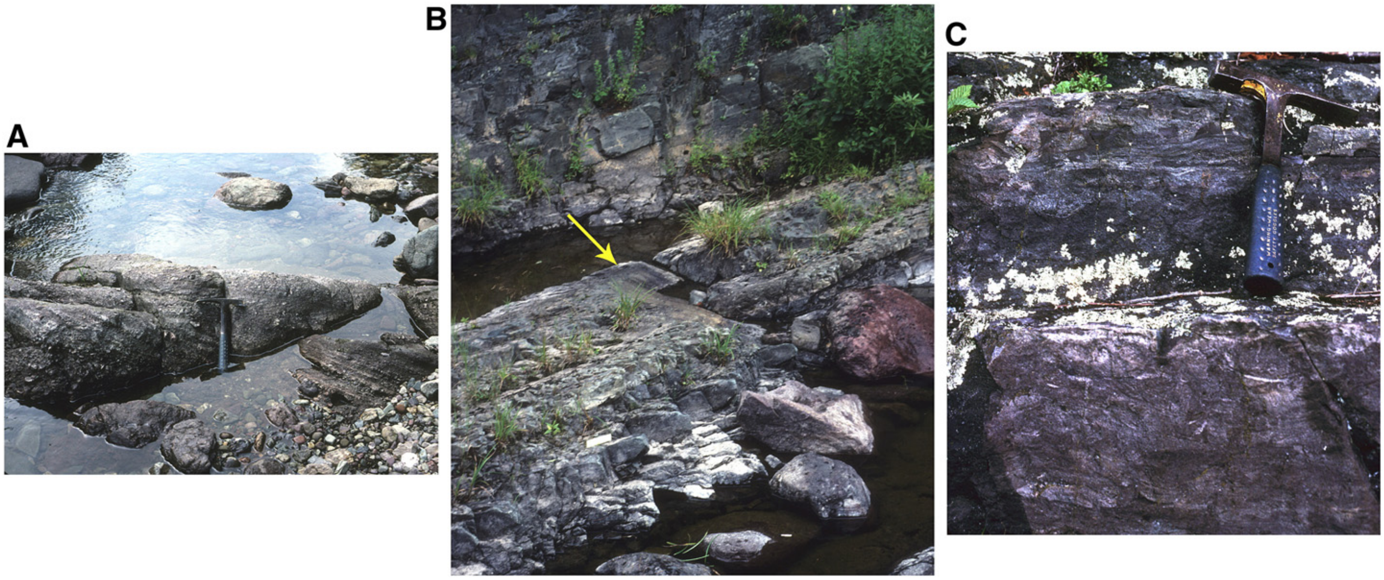

Fig. 5. Lithologies and depositional context of lower Mystery Lake Member diamictite in Hanford Brook East section. (A) Trough cross-bedded, braided stream deposits that underlie dropstone bed in lower Mystery Lake Member of the Chapel Island Formation; length of hammer 30 cm. (B) Tip of yellow arrow marks top of dropstone bed in lower Mystery Lake Member. (C) Hammer (30 cm long) is propped on dropstone bed; note outlines of small, teardrop-shaped ball-and-pillow structures.

Strata above the lower olive green mudstone include an alternation of mudstone and fine- to medium-grained, thin-bedded sandstone with increased burrow intensity higher in the section (Fig. 4). Thin, rusty-weathering, calcareous nodules at 191 m have a low diversity fauna of shelled metazoans of the *Watsonella crosbyi* Zone (Landing, 2004). These microfossils include the agglutinated foraminiferan *Platysolenites antiquissimus* Eichwald (1860)—which also occurs at 193.75 m (Fig. 4). The *W. crosbyi* Zone assemblage is the oldest diverse assemblage of skeletalized metazoans known in Avalon, and represents an important interval in the Cambrian evolutionary radiation (Landing et al., 1989; Landing and Westrop, 2004).

Detailed summaries of the lithology of the higher Mystery Lake Member at Hanford Brook East, Ratcliffe Brook, and in Saint John, New Brunswick are in Landing and Westrop (1998) and Landing (2004). The higher strata of the Mystery Lake consist of wave-dominated sandstone and minor mudstone (e.g., Landing and Westrop, 1998).

3.4. Diamictite bed

Interval 173.95–174.35 m is a diamictite bed with oversized clasts in a muddy matrix near the base of the olive green mudstone (Figs. 4 and 5B). Numerous slabs were cut through the bed, and the cut surfaces in Fig. 6 are representative of the bed's lithology. The dark olive green color of the diamictite's muddy matrix is identical to that of the thin interval of underlying mudstone (5 cm thick) that directly overlies the fluvial deposits, as well as that of the overlying mudstone (174.35–177.75 m).

Small areas of Fig. 6A (lower right corner) and Fig. 6B (lower left corner) illustrate the underlying olive green mudstone with its lenticular, glauconitic(?) quartz sandstone. This underlying mudstone is important in showing that standing water capable of floating ice abruptly followed deposition of the fluvial conglomerate and sandstone.

A variety of clasts occur in the diamictite. These include 1–3 cm-diameter pieces of brownish-green, fine-grained, structureless (i.e., presumably burrow-homogenized sandstone; Fig. 6A, lower central part of figure). This fine-grained sandstone lithology is minor in the Quaco Road Member, does not occur in strata that are vertically adjacent to the diamictite bed, and is prominent only in much higher strata of the Mystery Lake Member at Hanford Brook East (ca. 242–249 m).

In addition, isolated clasts up to 12 cm in diameter are present. They include terminal Cryogenian–Lower Ediacaran Coldbrook Group basalt; middle Proterozoic Greenhead Group, the only unit in southern New Brunswick with abundant, coarse-grained marble (see Currie, 1987); hydrothermal quartz; and feldspathic sandstone (Fig. 6B, left and centre of figure). The large marble clasts are unique to the diamictite, and were not found in the underlying fluvial deposits. The closest outcrops of Greenhead Group marble lie 20 km to the east of the Hanford Brook East section (Hayes and Howell, 1937, pl. 9 map), represent a unique through relatively rare lithology in the diamictite bed, and are the largest clasts (reaching 12 cm diameters).

As the underlying fluvial deposits show an almost due north transport direction, and Green Head Group marble is unknown south of Hanford Brook East, the diamictite bed cannot represent a reworking of the fluvial deposits. Silty mudstone laminae within the diamictite are compacted over the large clasts (Fig. 6A, note 1.5 cm, sandy mudstone lamina bent over the large marble clast). More importantly for the interpretation of the diamictite is the fact that the laminated silty mudstone immediately above some of the large clasts is disrupted, and this suggests that the large clasts penetrated the mudstone lamina before compaction (Fig. 6A, note outlined area marked “d.l.,” for disrupted laminae).

Characteristic features of the diamictite bed are somewhat teardrop-shaped, coarse-grained, feldspathic sandstone lenses up to 5 cm in diameter in a muddy matrix (Fig. 5C). These small, circular (as demonstrated by slabbed sections) lenses consist of coarse-grained sandstone in which the laminae are bent upward at the margin of the lenses. They are not comparable to the elongate spreite of teichichnid trace fossils, and are ball-and-pillow structures that resulted from the vertical sinking of sand lenses through the unconsolidated diamictite (Fig. 6B, symbol “b.p.”). These ball-and-pillow structures are interpreted to be the remnants of a feldspathic sandstone deposited within the olive green mudstone.

3.5. Chapel Island Formation diamictite interpreted as dropstone bed

As discussed in many reports (e.g. Mackiewicz et al., 1984; Benn and Evans, 1998), ice-rafted debris may be hard to distinguish from outsized clasts deposited by turbidity currents, debris flows, or grain flows. However, the diamictite shows no features consistent with the

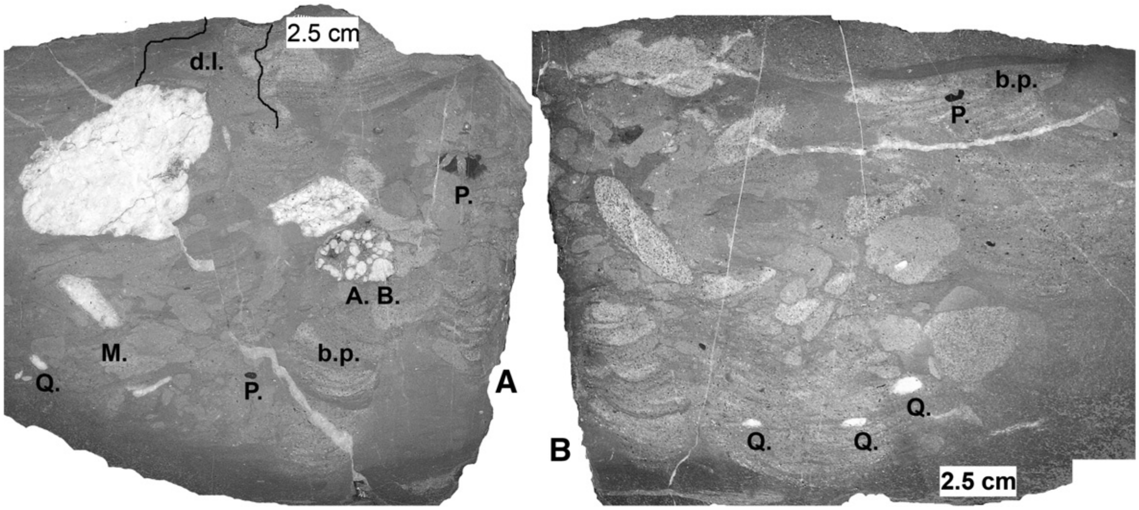

Fig. 6. Dropstone conglomerate slabs (specimen NBMR 1788), dark olive green colored. (A) Muddy matrix-dominated facies with abundant dark olive green mudstone clasts in centre of figure; note ca. 1.5 cm mudstone lamina bent over largest white marble clast. (B) Sand-rich matrix-dominated facies, note feldspathic sandstone clasts in centre and left side of slab. Abbreviations: A.B., amygdaloidal basalt from early Ediacaran (ca. 600 Ma) Coldbrook Group; b.p., example of small ball-and-pillow structures; d.l., disrupted laminae; M., Marble, middle Proterozoic Greenhead Group; P., phosphorite; Q., hydrothermal quartz clasts.

three latter mechanisms [e.g. no normal or inverse grading or inclined clasts that are stacked, or shingled on each other (Larsen and Steel, 1978; Nemec and Steel, 1984; Benn and Evans, 1998)]. The bed's muddy matrix precludes interpretation as a grain flow, as grain flows move by dispersive force between clasts and the muddy matrix would have damped movement as a grain flow. The diamictite does include features consistent with deposition by debris flow, including clasts suspended in a muddy matrix and a lack of lamination. However, a dropstone conglomerate interpretation is consistent with the disruption of the silty laminae above the large marble clast, the subsequent compaction of the broken edges of these laminae over the large clast, and the absence of any shear and rotation of the ball-and-pillow structures during debris flow motion. Finally, an interpretation of the diamictite bed as a transgressive lag at the base of the olive green mudstone can be dismissed because the diamictite lies within the mudstone, not directly at its base, and has an unwinnowed, muddy matrix.

As discussed by Thomas and Connell (1985), the presence of oversized clasts in a bed with non-erosive lower contacts (the diamictite lies within mudstone and 5 cm above the top of the fluvial deposits) and the draping of overlying sediments over the clasts are indicators of ice-rafted, dropstone conglomerates. Similarly, the lack of any evidence of rotation of the large clasts and the ball-and-pillow structures or the stretching out of the ball-and-pillow structures rules against deposition of the diamictite as a lodgment till by subglacial shearing (Alley, 2000; Van der Wateran et al., 2000) or as a debris flow. Production of the diamictite bed as the consequence of a volcanic eruption is not feasible, as volcanic clasts are relatively rare in the diamictite bed, fine-grained ash is not present in the bed, and sand-sized ash is not present in the feldspathic quartz arenite that forms the ball-and-pillow structures.

Numerous slabs were cut through the diamictite bed to examine its internal construction and the distribution and shape of its clasts. Slabbing was done as the bed is strongly indurated, and it is impossible to break the relatively soft and fragile marble and amygdaloidal basalt clasts free of the matrix. Fig. 6A is particularly representative of the shapes of the oversized clasts.

As summarized by Flint (1971, p. 165–169), although most glacial clasts have an inherited shape, clasts with a roughly triangular or pentagonal outline are relatively common in glacial deposits, and the flat faces are described as "faceted." In addition, some glacial clasts are angular and bound by fracture surfaces resulting from glacial detachment and crushing. Fig. 6A shows the "exotic" Proterozoic clasts in the diamictite bed—large marble clasts often with a roughly triangular (faceted) outline, while other clasts, as the small marble pebble and the amygdaloidal basalt pebble to the lower right of the large marble clast, have pronounced angular shapes. Striations, particularly on faceted faces, would be a strong indicator of glacial origin. However, as noted by Flint (1971), only a small percentage of clasts in till have striae. Indeed, brittle, coarsely textured clasts, such as the marble and amygdaloidal basalt in the diamictite bed, are more likely to be fractured than abraded during transport by glacial ice. As the Proterozoic clasts cannot be broken free of the diamictite bed and weather more rapidly on outcrop surfaces than the surrounding matrix, striae were not observed.

4. Biostratigraphy of the Chapel Island Formation dropstone bed

All of the Chapel Island Formation is referable to the Early Cambrian. Trace fossils in the lower and middle Quaco Road Member (Matthew, 1889; Hofmann and Patel, 1989) are consistent with a general Lower Cambrian correlation, and are younger than the *Trichophycus pedum* Zone ichnofossil assemblage that defines the global standard for the base of the Cambrian (Landing, 1994a). The significantly younger *Watsonella crosbyi* Zone of the lower Mystery Lake Member on Hanford Brook corresponds to an interval of small body fossils reported by Matthew (1889), and includes taxa (a gastropod, an operculate annelid, an orthothecid hyolith, an agglutinated foraminiferan, and a tommotiid) that indicate the lower, sub-trilobitic Lower Cambrian (Landing, 2004).

A conclusion that the lower Mystery Lake Member and its dropstone bed is earliest Cambrian is consistent with a 530.7 Ma date on the upper Mystery Lake in Saint John, New Brunswick (Fig. 2, locality SoS; Isachsen et al., 1994). U–Pb zircon geochronology and correlation by carbon isotope excursions show that the 530.7 Ma date on the upper Mystery Lake correlates into the sub-trilobitic Manykaian/Nemakit–Daldynian of Siberia (Maloof et al., 2005). In short, the dropstone conglomerate is referable to the Terreneuvian, the lowest series of the Cambrian (Landing et al., 2007). *Watsonella crosbyi* Zone fossils occur above the depositional sequence boundary in the Chapel Island Formation in southern New Brunswick. However, the zone's base occurs below and persists through the sequence boundary in east Newfoundland (Landing, 1989; Landing et al., 1989). A diachronous lowest occurrence of *W. crosbyi* Zone fossils between eastern Newfoundland and southern New Brunswick reflects taphonomic control—with suitable mudstone facies with minor carbonates and local pyritization of the fossils occurring below the sequence boundary in the upper Quaco Road Member in Newfoundland (Landing, 1989; Landing et al., 1989). In southern New Brunswick, carbonate nodules with pyritized *W. crosbyi* Zone fossils (now weathered to hematite) appeared only in the lower Mystery Lake Member (Landing, 2004).

The *Watsonella crosbyi* Zone forms the middle Terreneuvian Series in eastern Newfoundland (Landing et al., 1989; Landing and Westrop, 2004). Li et al. (2007) proposed the *W. crosbyi* Zone base as a potential base of the Terreneuvian Series' second stage. Although the *W. crosbyi* Zone base is diachronous in Avalon, Li et al.'s (2007) suggestion would place the dropstone bed and the Quaco Road–Mystery Lake sequence boundary in the boundary interval between the upper Fortunian Stage (lower Terreneuvian) and the presently unnamed second (and upper) stage of the Terreneuvian.

5. Geochronological, eustatic, and carbon isotope significance of a middle Terreneuvian glaciation

The upper Quaco Road Member of the Chapel Island Formation in Avalonian New Brunswick is a shoaling-up succession that ends at an eroded type 1 depositional sequence boundary (i.e. Van Wagoner et al., 1988). By comparison, the lower Mystery Lake Member is a transgressive succession that features a physically abrupt transition from lowstand, braided stream deposits into overlying marine-deposited mudstone and minor sandstone lenses. Lying in the lowest part of the marine deposits is a dropstone conglomerate. This bed suggests the melting of sediment-charged ice and the release of large clasts and abundant sand into mud that overlay the fluvial sandstone and conglomerate of the basal Mystery Lake Member.

A transgressive package from braided stream deposits into overlying marine mudstone with a dropstone bed all suggest that the lowest Mystery Lake Member records a sea-level rise temporally associated with deglaciation. An approximately coeval sequence boundary within the *Watsonella crosbyi* Zone is also recognized between the Quaco Road and Mystery Lake Members approximately 650 km to the northeast in the southern Burin Peninsula, eastern Newfoundland. In this latter area, the sequence boundary lies at a contact between a 60 cm-thick interval of condensed fossil hash and mud mound limestones at the base of the Mystery Lake Member and an underlying paleosol with caliche and peds developed in red mudstones at the top of the Quaco Road Member (Landing et al., 1989, fig. 4.1; Landing, 2004). These southern New Brunswick and eastern Newfoundland localities lie within the Avalon terrane, and an apparently coeval sea-level fall–rise couplet could be epeirogenic. However, the association of relative sea-level rise with evidence of deglaciation at localities separated by ca. 650 km away suggests a eustatic event.

Major eustatic events have been associated with stable carbon isotope excursions—perhaps as a result of the burial of light carbon with sea-level rises and the expansion of anoxic zones (e.g. van Houten and Arthur, 1989). If Avalon and related regions underwent a significant glaciation and subsequent deglaciation within the earliest Cambrian, the eustatic rise and transgression could have led to significant carbon burial and development of a strong negative $\delta^{13}C$ excursion.

Interestingly, Zhu et al. (2006) documented two presently unnamed negative excursions in southern China. These excursions occur between the basal Cambrian (BACE), i.e. basal Terreneuvian, excursion (Landing et al., 2007), and the strong negative excursion (ZHUCE) that Zhu et al. (2006) regarded as corresponding to the extinction of many small shelly fossils in Stage 2 of the upper Terreneuvian Series.

As discussed above, the *Watsonella crosbyi* Zone and dropstone bed can be regarded as lying in the Fortunian Stage–Stage 2 transition of the Terreneuvian Series. This transitional interval corresponds to a strong, presently unnamed negative $\delta^{13}C$ excursion in the Zhu et al. (2006, fig. 1) carbon excursion diagram for the Cambrian, and thus to a possible strong (glacio)eustatic rise.

Maloof et al. (2005, fig. 8) recorded two strong negative $\delta^{13}C$ excursions in the earliest Cambrian of southern Morocco. The stronger and younger of these is in evaporitic dolostone of the Adoudou Formation, the oldest cover unit on the Pan-African orogen, and is calculated to be ca. 529 Ma in age. This excursion is correlated with the upper part of the significantly sub-trilobitic Manykaian/Nemakit–Daldynian Stage of Siberia. This upper, negative $\delta^{13}C$ excursion, even with uncertainties in geochronology and correlation, is apparently younger than a 530.7 Ma date on the uppermost Mystery Lake Member. A lower, negative $\delta^{13}C$ excursion in the Adoudou that Maloof et al. (2005) place at ca. 535 Ma and correlate with the middle Manykaian/Nemakit–Daldynian may correspond to the sea-level rise recorded by the lower Mystery Lake Member in southern New Brunswick and eastern Newfoundland, and the possible melting of glacial ice in Avalon.

6. Potential supporting evidence from the Cambrian of Avalonian Ireland

Study of Lower Palaeozoic successions from Rhode Island to Wales (e.g. Landing, 1996a) has not revealed more evidence of Cambrian glaciation on the Avalon platform. However, evidence of Cambrian glaciation should be preserved as dropstones and other features in the deep-water, continental slope–rise successions (e.g. Gander and Meguma Groups) northwest and southeast (in modern coordinates) of Avalon. Indeed, potential dropstones are now known in the deep-water, Avalon terrane of southeast Ireland.

In southeast Ireland, the Cambrian is represented primarily by the Booley Bay Formation—largely rhythmically thinly interbedded siltstones and mudstones with local, poorly sorted sandstone beds and thick mudstone units. This succession is interpreted as a deep-marine slope sequence of contourite-modified turbidites with local debris flows. Slow-downs in deposition are marked by hemipelagic sedimentary rocks (MacGabhann, 2007b). The age of the Booley Bay Formation has been difficult to ascertain, with palynological evidence providing the sole indication of age. However, even existing palynological studies have produced equivocal results—with the succession variously considered Lower Cambrian to Lower Ordovician (Gardiner and Vanguestaine, 1971), Middle Cambrian (Smith, 1981), Upper Cambrian (Moczydłowska and Crimes, 1995), or Middle and Upper Cambrian (Vanguestaine and Brück, 2008). However, the lack of ichnofossil diversity, along with the absence of skeletal fossils, the presence of Ediacaran-like soft-bodied fossils, and the thickness of the succession (MacGabhann, 2007b; MacGabhann et al., 2007) suggest the Booley Bay is Lower–Middle Cambrian.

In this context, the presence of isolated outsized clasts (Fig. 7) in siltstone beds in the Booley Bay Formation may be supportive of the conclusion that there was Early Cambrian glacial activity in Avalon. The clasts occur in at least four siltstone beds in the Booley Bay Formation at Grange Strand, Fethard-on-sea, Co. Wexford, in low outcrops covered by high tide (and intermittently by beach sand). The clasts range in size from 45–140 mm, and are moderately angular to well rounded, with variable degrees of sphericity. While there is no evidence to suggest an extrabasinal origin, they are coarser grained than the containing sediment, and often protrude from erosional surfaces (Fig. 7A).

Deep-sea siderite nodules are present in the overlying Ribband Group (Shannon, 1977). However, the Booley Bay Formation clasts are quite different to these nodules, and do not react with heated acid; thus, a concretionary interpretation is not considered likely. Indeed, the variation in size, shape, and rounding of the clasts implies sedimentary transport of various degrees. However, while debris flows and turbidity currents were undoubtedly operating in the region, it is unlikely that they were responsible for these outsized

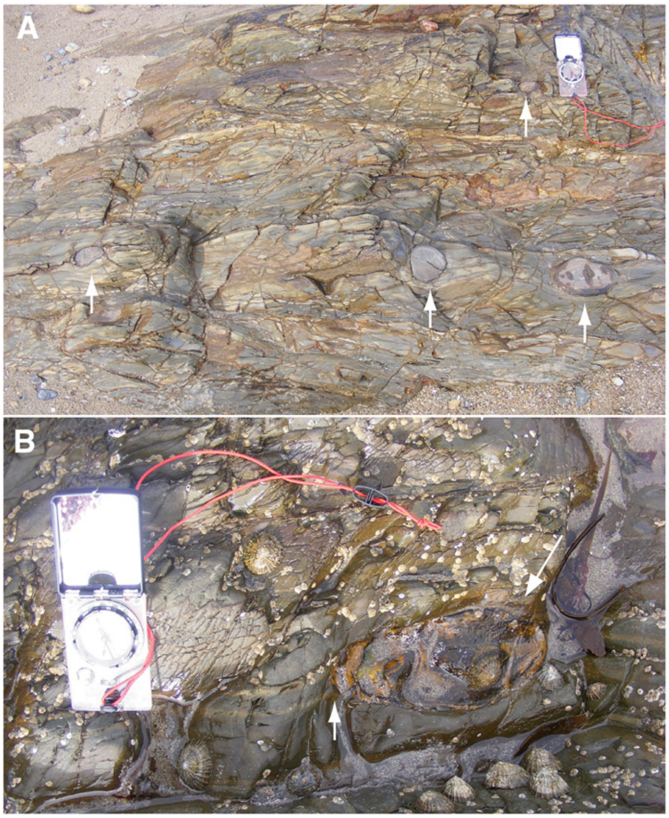

Fig. 7. Outsized clasts in the Booley Bay Formation at Grange Strand, Fethard-on-sea, southeast Ireland. Younging direction toward top of the images. (A) Four outsized clasts in two beds (arrows); middle clast in the bottom layer has a furrow which may represent glacial striation. Irish Grid Reference S80062 05452. (B) Outsized clast which appears to cut sedimentary laminations (left arrow; right arrow shows upper right corner of clast). Irish Grid Reference S80070 05504. Opened transit measures 20 cm.

clasts as the debris flow units are highly recognizable, and the turbidity currents were without doubt distal and unable to carry sediment above silt-size. Furthermore, clasts in debris flows elsewhere in the formation are uniformly cohesive, rolled-up fragments of local bedding, and clearly distinct from these outsized clasts. No other rounded clasts have been observed in the Booley Bay Formation (MacGabhann, 2007b).

One of the clasts is observed to have a furrow which may represent a glacial striation, and laminations are also observed to be deformed below, and in one case cut by, these clasts (Fig. 7B). Thus a glacial dropstone interpretation is highly plausible. It may also be postulated that the protracted slow-downs in deposition that are represented by metre-thick, organic mudstone hemipelagite units are consistent with an interpretation of glacial activity, as these could potentially result from glaciation and an interruption of terrestrial sediment supply to the basin with the freezing of ancient rivers and streams.

Only a far more precise dating of the Booley Bay Formation can confirm a temporal link between these outsized clasts and those in New Brunswick. Unfortunately, the section in which they occur is heavily deformed with a significant amount of faulting, and the beds cannot be traced for more than a few metres laterally, making both their position within the formation and their age almost impossible to determine. However, the evidence from the Booley Bay Formation is nonetheless supportive of the hypothesis that Avalon was subject to glacial activity in the Cambrian.

7. Palaeogeographic significance of Avalonian Cambrian glaciation

Minor variants on palaeogeographic maps that show the Avalon terrane as allied to the West African margin of Gondwana through the Cambrian regularly appear in the literature (e.g. Scotese and McKerrow, 1990; McKerrow et al., 1992; Fortey and Cocks, 1992, 2003; Álvaro et al., 2003; Linnemann et al., 2008; Palacios et al., 2008). This palaeogeographic reconstruction actually had its genesis in a continental drift proposal by Choubert (1935), who noted that a number of late Early–Middle Cambrian trilobite genera are common to Morocco and to such Avalonian areas as Wales and eastern Newfoundland, and concluded that these regions were adjacent in the Cambrian. More recently, precisely the same approach was used, for example, by Álvaro et al. (2003, fig. 5) to show that "Early" Cambrian trilobite faunas of Morocco and Avalon are similar, and this means that Avalon existed as a sort of appendage attached to the West African margin of Gondwana through the Cambrian (Álvaro et al., 2003, fig. 3).

The problem common to all of these palaeogeographic maps is that they are based on late Early Cambrian trilobites [ca. 511 Ma., Landing et al., 1998] and younger Cambrian faunas. Unfortunately, these palaeogeographic reconstructions have not considered older Cambrian faunas or the differences between the major Cambrian lithofacies associations of Avalon and West Gondwana.

As noted by Landing (2005), Pillola's (1993, fig. 4) map of the distribution of the oldest Gondwanan trilobites provides a useful cartoon that can be used to debunk the Early Cambrian unity of Avalon and Gondwana. Pillola (1993, fig. 4) shows that none of the earliest trilobites (fallotaspids and bigotinids) of Gondwana (Morocco, Iberia, France, Sardinia) occur in supposedly "adjacent" Avalon (which has olenellids not present in "adjacent" Gondwana). Similarly, West Gondwana has diverse archaeocyathans and their build-ups, but archaeocyathans remain unknown in Avalon.

The remaining arguments against an Avalon–West Gondwana unity in the Cambrian are lithostratigraphic and climatic, and in the fact that evidence exists for the palaeogeographic isolation of Avalon by the earliest Cambrian (Fig. 1A). Thick (up to 2.2 km), tropical carbonate platform successions began accumulating in the latest Ediacaran or earliest Cambrian, and comprise the oldest Pan-African and Cadomian cover successions in southern Morocco and Normandy (Doré, 1969; Landing 1996a, 2005; Geyer and Landing, 2006). By comparison, coeval Avalonian successions are invariably siliciclastic-dominated, with cold-water, fossil-hash limestone beds that accumulated near the shoreline of shale-dominated basins. These limestone beds are generally much less than a metre thick, and never show the ooids, thrombolites, and evaporitic minerals that characterize parts of the Gondwanan carbonate successions (Myrow and Landing, 1992; Geyer and Landing, 1995; Landing and Westrop, 1998, 2004; Landing, 2005; Geyer and Landing, 2006).

Finally, the development and location of thick Cambrian–Ordovician continental slope and rise prisms emphasize the separation of Avalon from all other Early Palaeozoic paleocontinents (Landing, 2005). These deep-water successions include the Gander Group–Wexford terrane (southeast Ireland with the Booley Bay Formation)–Lake District (northern England) and Meguma Group–Brabant–Rhenohercynian anticlinoria on the northwest and southeast margins of Avalon, respectively. These deep-water successions show that Avalon was an insular microcontinent isolated from other palaeocontinents by the Cambrian (e.g. Keppie et al., 2003; Landing, 2005; Fig. 1A, B).

Dropstone conglomerates in New Brunswick and Ireland are consistent with a geographic separation between high south latitude Avalon and tropical West Gondwana by the earliest Cambrian. The latitudinal separation and climatic distinctiveness of the continents followed from the rifting of Avalon and its movement along transform faults away from Gondwana (Keppie et al., 2003). This tectonic activity began in the Ediacaran, and not as late as the Early Ordovician (fide Fortey and Cocks, 1992, 2003) (Fig. 1).

As the Cambrian progressed, Gondwana also began a southerly movement into temperate latitudes (Theokritoff, 1979; Burrett et al., 1990; Landing, 1996a), which led to the loss of significant carbonate deposition in most areas of Gondwana as it crossed into a cooler climate regime. The persistence of carbonate platform deposition (Láncara and Vegadeo Formations) into the middle Middle Cambrian of the Cantabrian and West Asturian–Leonese Zones/terranes suggests that northern Iberia was the last West Gondwanan region to pass south and out of tropical latitudes (Fig. 1).

The break-down of provincial barriers and the sharing of trilobite genera took place beginning in the late Early Cambrian as the southern movement of West Gondwana and the northerly movement of Avalon brought them into comparable latitudes and climate belts. By the Ordovician–Silurian boundary interval, West Gondwana lay at the South Pole and had extensive continental glaciation, while carbonate platform facies and coral reefs appeared in the late Early Silurian of equatorial Avalon.

8. Concluding remarks

Dropstones in Avalonian successions at the Hanford Brook East section, southern New Brunswick, and in the Booley Bay Formation, southeast Ireland, provide the first evidence for glaciation in the Cambrian. It would be particularly persuasive to find a record of thick till deposits and glacially deformed glacial sedimentary rocks deposited on striated bedrock surfaces as in the Late Ordovician of north Africa (e.g. Le Heron, 2007). However, the earliest Cambrian section on Hanford Brook East does indeed have a thin diamictite bed with evidence that its Proterozoic clasts are dropstones that cut now overlying silty mudstone laminae that are draped over the clasts. Furthermore, these clasts occur within a marine mudstone and lie above a transition from underlying fluvial deposits. Finally, the clasts have the shapes (triangular and strongly angular) associated with glacially modified stones. The relatively minor evidence of glaciation at the Hanford Brook East section may simply reflect the section's location near the margin of the extent of Early Cambrian glaciation in Avalon. Similarly, dropstones in the Booley Bay Formation, southeast Ireland show the presence of sediment-charged floating ice outboard and northwest of the Avalon platform, and provide further evidence of Cambrian glaciation.

Despite a thorough search through Chapel Island Formation deposits on the marginal Avalonian platform of southern New Brunswick, central Cape Breton Island, and eastern Newfoundland, no additional evidence for glacial deposits has yet been found, although all these areas show evidence for a sea-level fall–rise couplet for 650 km along the Avalonian marginal platform in the middle Chapel Island (Landing, 2004). The only other area on the marginal platform that has strata equivalent to the Chapel Island Formation is North Wales, where the “Basal Series” and “Purple, red, and green slates” below the Dorothea and Red Grits are lithologically comparable to and are correlated with the Rencontre and Chapel Island Formations, respectively (Landing, 1996a). It is possible that the middle of the “Purple, red, and green slates” may show evidence of Early Cambrian glaciation, although inspection of this poorly exposed succession in North Wales (e.g. Wood, 1969) by E. Landing and P. Myrow in 1994 did not lead to recognition of any potentially glacially deposited sedimentary rocks. Although relatively thorough study of the available outcrops seem to militate against additional evidence for an Early Cambrian glaciation on the Avalon platform, the recognition of dropstones in southeast Ireland suggest that more significant Cambrian glaciomarine deposits may occur in the Gander and Meguma successions marginal to Avalon (Fig. 1A).

If the proposed glaciation–deglaciation couplet in Avalon had an effect on eustatic levels, then a shoaling–possible subaerially eroded type 1 sequence boundary–transgression succession may occur in the middle Adoudou Formation in Morocco and Manykaian/Nemakit–Daldynian of Siberia (e.g. Maloof et al., 2005). However, evidence for such a eustatic history may be lost at the apparent long hiatus on many parts of the South China Platform between the oldest small shelly fauna interval (traditionally “Meishucunian A”) and the much younger, probably Tommotian-equivalent “Meishucunian B” small shelly faunas (Landing, 1994a,b), the latter of which occurs in an interval with a strong positive δ^{13}C excursion (Brasier et al., 1990) and cannot correlate with the *Watsonella crosbyi* Zone.

The record for a ca. 535 Ma glaciation–deglaciation interval in the *Watsonella crosbyi* Zone includes shoaling through the upper Quaco Road Member of the Chapel Island Formation followed by development of an erosive sequence boundary that extends for approximately 650 km along the marginal Avalonian platform. Deglaciation is recorded by a transgressive succession (braided stream deposits and overlying marine mudstone) of the overlying Mystery Lake Member. Development of the dropstone bed in the lower Mystery Lake Member near the base of the marine mudstone records the melting of sediment-charged ice at sea level. Presence of this dropstone bed emphasizes that Avalon separated (probably from Gondwana) by the very earliest Cambrian or Ediacaran, and that it lay at a high south latitude while West Gondwana was tropical. Palaeogeographic maps that show a unity of Avalon and Gondwana through the Early Ordovician are in error because the data shows fundamental differences in cover succession lithology through the Early Cambrian, in the composition of the earliest trilobite assemblages, and in the absence of archaeocyathans in Avalon.

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