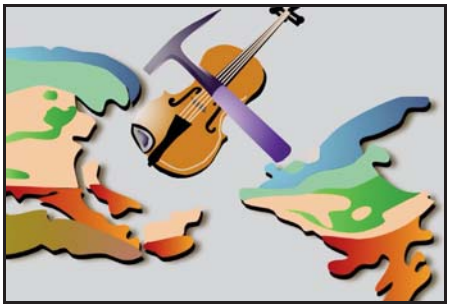
HAROLD WILLIAMS SERIES



Ode to Field Geology of Williams: Fleur de Lys Nectar Still Fermenting on Belle Isle

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SUMMARY
Throughout the 1960s, Hank Williams put Newfoundland on the proverbial global map as one of the most complete cross-sections of the Appalachian Orogen, and he became a champion attractor to this unique geological laboratory. By the end of the 1960s, Williams, together with Bob Stevens, had mapped the rocks of Belle Isle in the treacherous waters north of the Long Range Peninsula, and suggested their siliciclastic rocks were equivalent to those of the Fleur de Lys type sections on the Burlington Peninsula some 200 km away across White Bay, and by implication that the underlying Laurentian basement on Belle Isle should have its counterpart there too. New U–Pb geochronology on zircon from two samples of possible basement to the Fleur de Lys Supergroup is presented here. These data verify unequivocally the wisdom of the original suggestions based on dedicated field work. The new data also provide evidence that by the earliest Ordovician (ca. 483 Ma), high pressure-low temperature metamorphism at depths in excess of 30 km occurred in Fleur de Lys Supergroup domains. The tectonic implications of these findings are explored, and from this it emerges that only new mapping integrated with high-resolution geochronology *and* thermochronology are required, both on Belle Isle and in the Fleur de Lys Supergroup, to advance beyond the standards set by Hank Williams.

SOMMAIRE
Tout au long des années 1960, Hank Williams aura mis Terre-Neuve sur la carte mondiale comme étant le lieu de l'une des coupes géologiques les plus complètes de la chaîne de l'orogène des Appalaches, devenant ainsi un attracteur incontournable pour ce laboratoire géologique unique. À la fin des années 1960, Bob Stevens et lui avaient cartographié les rochers de Belle-Isle, dans les eaux traîtresses du nord de la péninsule de Long Range, et proposé que leurs roches siliciclastiques étaient l'équivalent de celles des sections des coupes types Fleur de Lys sur le péninsule Burlington, quelques 200 km au delà des eaux de White Bay, et en conséquence, que le sous-sol des Laurentides sous-jacent à Belle-Isle devrait avoir sa contrepartie là aussi. Une nouvelle datation géochronologique U-Pb sur zircon sur deux échantillons, potentiels du substratum du supergroupe de Fleur de Lys, est présentée ici. Ces données corroborent sans équivoque la sagesse des suggestions originales basées sur un travail de terrain méticuleux. Les nouvelles données fournissent aussi la preuve qu'au tout début de l'Ordovicien (env. 483 Ma), qu'un métamorphisme de haute pression et de basse température a sévi à des profondeurs de plus de 30 kilomètres dans les domaines du supergroupe de Fleur de Lys. Les implications tectoniques de ces résultats sont considérées, et de cela il se dégage que seule une nouvelle cartographie intégrée à une géochronologie et une thermochronologie de haute résolution, tant sur Belle-Isle et dans le supergroupe de Fleur de Lys, pourront permettre d'ajouter aux résultats obtenus par Hank Williams.

PRELUDE
By the early 1960s, Hank (Harold) Williams understood both the neglect and the potential attractor of the northern coast of Newfoundland as the best and most complete symmetrical cross-section of the Appalachian system (Williams 1964). In this classic paper, based on dedicated geological mapping that formed the foundations of all his work, Williams lamented that the two-sided nature of this mobile belt had not yet 'taken root among students of regional tectonism, particular-

Geoscience Canada, v. 41, http://dx.doi.org/10.12789/geocanj.2014.41.043

ly the proponents of continental accretion', despite the general knowledge among paleontologists by then about trilobite faunal provinciality in western and eastern Newfoundland, which are quite distinct. He stressed that such field observations contradicted the then prevailing view that the Appalachians in Newfoundland, and by inference elsewhere, simply formed by accretion of a Paleozoic welt along the southeast border of Laurentia in Paleozoic times. Central to Williams' deduction was that along both extremities of east and west Newfoundland, old continental crust was overlain by Lower Paleozoic sedimentary shelf sequences (east- and west-ward thinning clastic wedges) that imperceptibly merged with deep water sedimentary rocks and variable volcanic rocks associated with ultramafic intrusions comprising central Newfoundland. He knew this because he'd mapped it all: "*The Appalachian Paleozoic mobile belt in northeastern Newfoundland is symmetrical, bounded on the northwest and on the southeast by Precambrian rocks overlain by Lower Paleozoic shelf deposits*". He believed that possibly the oldest Precambrian rocks of Newfoundland constituted a seaward volcanic arc that formed the continental border and represented continental spreading in Late Precambrian time, inferring a *Late* Precambrian origin for the ocean, younger than tacitly assumed; and that the 'Paleozoic geosynclinal rocks accumulated towards the continental side of this feature'. Williams was adamant therefore that 'the geologic history of the Appalachians goes back far beyond the Paleozoic, and this early history must be included in any complete treatment of their evolution. Only when this is done will the regional tectonic significance of the system with respect to the rest of North America be fully appreciated'. With this vision, Williams ignited renewed excitement and stimulated a new era of re-mapping across Newfoundland.

I was still in high-school when Hank first synthesized his field observations in this manner, and I was certainly not able to appreciate until years later how these early seeds of geologic, reaped from his Appalachian field experiences in Newfoundland, would blossom into game-changers. How fast these seeds penetrated the minds of leading geo-scholars at a time that the foundation of plate tectonics was also emerging is difficult to reconstruct; but, within 5 years, testing Williams' ideas against the 'new global tectonic' interpretations of the Appalachians and related mobile belts under the watchful eye of Marshal Kay, was well on the way by a new cohort of field geologists that had feasted on the geo-brew of Williams — and that of his student Bob (R) Stevens.

I learned about some of this ferment in Newfoundland from Adrian Phillips, at Trinity College, Dublin; and during a memorable student field-trip in 1968 across the Irish Caledonides, led by John Dewey, both of whom had recently attended the famous 'Gander Conference' in Newfoundland (August 24–30, 1967) at which some of the new geotectonic ideas were expanded in debate and in its 'Guide to Field Trips'.

During that 1968 transect in the west of Ireland I became intrigued by the power of these new ways to interpret the field observations of Williams and his colleagues at the Geological Survey of Canada (e.g. Neale and Nash 1963). With hindsight, it seems legitimate then to speculate that had Hank Williams delayed his 1964 paper by only a few years during this time of profound changes in geology and the tectonic interpretations of old orogens, he would surely have inculcated in it the neo-modern ideas — for once a more robust geologic blueprint for Plate Tectonics and the Wilson Cycle (Wilson 1966) had emerged (1967–1968), it was a relatively simple step to interpret the Newfoundland sections in a framework of the emerging Plate Tectonic paradigm, and which subsequently culminated in a flurry of classic papers by John Dewey and Jack Bird on linking geology and plate tectonics of the Appalachians and Caledonides (Dewey 1969; Bird and Dewey 1970; Dewey and Bird 1971).

By 1969, the transition to a new geological order for the Appalachians was well on the way, and I became embroiled in it when inducted into John's Dewey field-team in 1969, as one of his PhD students. I arrived in Newfoundland with trepidation, which soon dissipated through warm and fluid welcomes that matched that of the Irish, and joined a group of compatriots to map where Williams and Stevens had left their marks.

At the same time Williams and Stevens continued unabated with their own field work; and in 1969, the year I started my first field season on the Fleur de Lys (FdL) rocks of the Burlington Peninsula, Hank and Bob published a paper on the geology of Belle Isle that caught my imagination (Williams and Stevens 1969; a heavily annotated copy of which I still have in my possession). During their coastal survey in the 'rather treacherous waters of the strait of Belle Isle', they mapped the unconformable relationship between Precambrian Grenville basement and the northern extremity of the sedimentary rocks of the Paleozoic Appalachian 'miogeosynclinal' belt, now better known as the Humber Zone (or Humber Margin; e.g. Williams 1979; van Staal and Barr 2012; van Staal et al. 2013).

Williams and Stevens speculated that volcano-sedimentary sequences on Belle Isle are the relatively pristine lithostratigraphic equivalents of the metamorphic Fleur de Lys Supergroup (FdLS) on the Burlington Peninsula, 150–200 km to the southeast across White Bay, where I was to spend the next 3 field seasons, after John Dewey had assured me this was a great place to investigate the internal part of the Humber Zone.

The Belle Isle paper proved an enduring inspiration: during my field work along eastern White Bay and its inlets, I continued to encounter FdL rock types remarkably similar to those mapped by Williams and Stevens from Belle Isle flanking western White Bay where they described gently deformed, fluvial conglomerate and shallow water coarse clastic sedimentary rocks (Bateau Formation), intruded by ca. 615 Ma mafic dykes feeding overlying tholeiitic basalt (Lighthouse Cove Formation), overlain by coarse arkosic siliciclastic rocks (Bradore[-like] Formation, but of unknown age), and in turn by upper Lower Cambrian shale and limestone (Forteau and White Point Formations). 'Seeing' through the metamorphic and deformation overprints on the FdL rocks then became a major challenge to test this hypothesis of Williams and Stevens.

I never saw John Dewey that first summer, as he was concentrating on the drafts of his now famous papers with Jack Bird: "*at the moment I am writing hard and....…..I may not get to see you this summer, although I shall be in western Newfoundland for a time. My schedule in Newfoundland is very tight – Jack Bird and I have a lot of work to accomplish in a very short time. Don't worry if I don't see you – just work hard and enjoy your rocks…*" (letter, April 1969). By the time I finally met up with John Dewey (and Jack Bird) for a few days during my last field season in 1971, I had benefited from an Appalachian transect with Hank Williams along the Trans-Canada Highway in Newfoundland; and from field-mentoring by a host of contemporary mappers: Ron Smyth, Stuart Sutton, Allan McCann and my other compatriots in John Dewey's team, John Bursnall and Bill Kidd. Armed with this help, Hank's memory bank of Newfoundland rocks, and Dewey's preprints, I was able to piece together a satisfactory story of the Fleur de Lys rocks, consistent with the observations by Williams and Stevens on Belle Isle: a late Neoproterozoic to early Paleozoic siliciclastic sequence unconformably overlying a Grenville-like basement. But not everyone was convinced.

Naturally since 1972, the geology of the western margin of Newfoundland has been refined by many superb geologists. One of the first to synthesize the growing knowledge over the following decade was Hibbard (1983). Subsequently others have taken this much further still, especially following integration with seismic profiles and ever more precise dating techniques (e.g. Williams 1979; Waldron et al. 1998; Skulski et al. 2010; van Staal and Barr 2012; van Staal et al. 2009a, b 2013). But what have remained undetermined are the age and even the very existence of possible basement to the Fleur de Lys sedimentary rocks of the Burlington Peninsula (e.g. Piasecki 1987, 1988). Hibbard (1983), amongst others, remained undecided and 'lumped' the rocks of the western Burlington Peninsula into a relatively conformable Fleur de Lys Supergroup sequence, including 'my basement' rocks, which he renamed the East Pond Metamorphic suite (EPMS), and subsequent workers appear to have avoided testing the wisdom of this.

In a way then, this contribution with my long-standing friend and peer Richard Armstrong, reflects a small pleasurable trip along memory lane, back to that exciting time in Newfoundland, to pay tribute to the field mapping legacy of a giant in the geological history of Canada; and to finally provide a test of the 1969 paper by Williams and Stevens that has episodically taunted me ever since completing my PhD: wherever I re-located over the past 40-odd years, I kept two key samples of Fleur de Lys 'basement' rocks amongst my belongings, ready for this test. Here we present some modern U–Pb analyses on zircon from those two samples to satisfy my curiosity of unfinished geo-business.

FLEUR DE LYS (FdL)

In 1941, James Osborne Fuller first used the term Fleur de Lys for a series of rocks exposed along the eastern headland at White Bay, following his field work there in 1937. These rocks have since been inculcated to define the western metamorphic zone of the Central Mobile Belt of the Appalachian fold belt in Newfoundland exposed on the Burlington Peninsula, along the eastern edge of White Bay and inland farther to the southwest (Baird 1951; Neale and Nash 1963; Neale and Kennedy 1967; Williams 1969; Church 1969; Kennedy 1971; Hibbard 1983; Cawood and Nemchin 2001; van Staal et al. 2013; Fig. 1). The FdL sequence was elevated to Fleur de Lys Supergroup (FdLS) by Church in 1969 (see Hibbard 1983, for an excellent detailed historical account), and is interpreted as a deformed late Neoproterozoic to early Paleozoic clastic sequence deposited along a passive continent–ocean transition zone flanking the Iapetus Ocean and its stable Laurentian continental shelf of Mesoproterozoic (Grenville-age) basement. This continental foundation is well exposed to the west of White Bay as the Long Range Inlier, the northeasternmost basement inlier of the Appalachian Orogen (Williams 1969; Bird and Dewey 1970; Bursnall and de Wit 1975; Hibbard 1983; Waldron et al. 1998; Cawood and Nemchin 2001; Heaman et al. 2002; van Staal et al. 2009a, 2013; van Staal and Barr 2012).

The Laurentian continental margin was initiated during a long period of rifting, characterized by mafic magmatism and coarse siliciclastic sediments deposited during the late Neoproterozoic between ca. 620–550 Ma (lower FdLS) before and during opening of the Iapetus Ocean. Thereafter, thermal subsidence created room for a transgressive Lower Cambrian sequence, generally interpreted to represent a rift–drift transition (upper FdLS; 540–530 Ma). Subsequently, the Laurentian passive margin was tectonically 'drowned' as a foreland basin in response to its collision with outboard arc-related Iapetus oceanic lithosphere. This resulted in the cessation of carbonate sedimentation on the continental shelf and the input of clastic detritus from the central Appalachian orogen (Williams and Stevens 1969, 1974; Stevens 1970; Bird and Dewey 1970; Dewey and Bird 1971; Strong and Williams 1972; Cawood and Nemchin 2001; van Staal et al. 2009a, 2013, and references therein).

In more detail, the Laurentian Margin in Newfoundland, also known as the Humber Zone (Fig. 1) underwent multiple episodes of rifting, as evidenced by rift-related magmatism between 615 and 580 Ma and between 565 and 550 Ma, in the process opening Iapetus and a small early Ordovician seaway, the Taconic seaway (e.g. Cawood et al. 2001; Waldron and van Staal 2001; van Staal et al. 2013). This resulted in an outer passive Humber Zone of ultra-thinned continental crust with mafic volcanic rocks directly underlain by old continental mantle lithosphere, similar to seaward-dipping volcanic rocks and continental mantle material at the Continent-Ocean Boundary (COB) zones of the North and South Atlantic margins (e.g. van Staal and Barr 2012; van Staal et al. 2013). This COB transition between the Humber and Dunnage Zones is preserved as a complex tectono-stratigraphic sequence known as the Birchy Complex (Bursnall 1975; Hibbard 1983; Skulski et al. 2010; van Staal et al. 2013; Fig. 1).

Outboard of the Humber Zone, the FdL is flanked by an amalga-

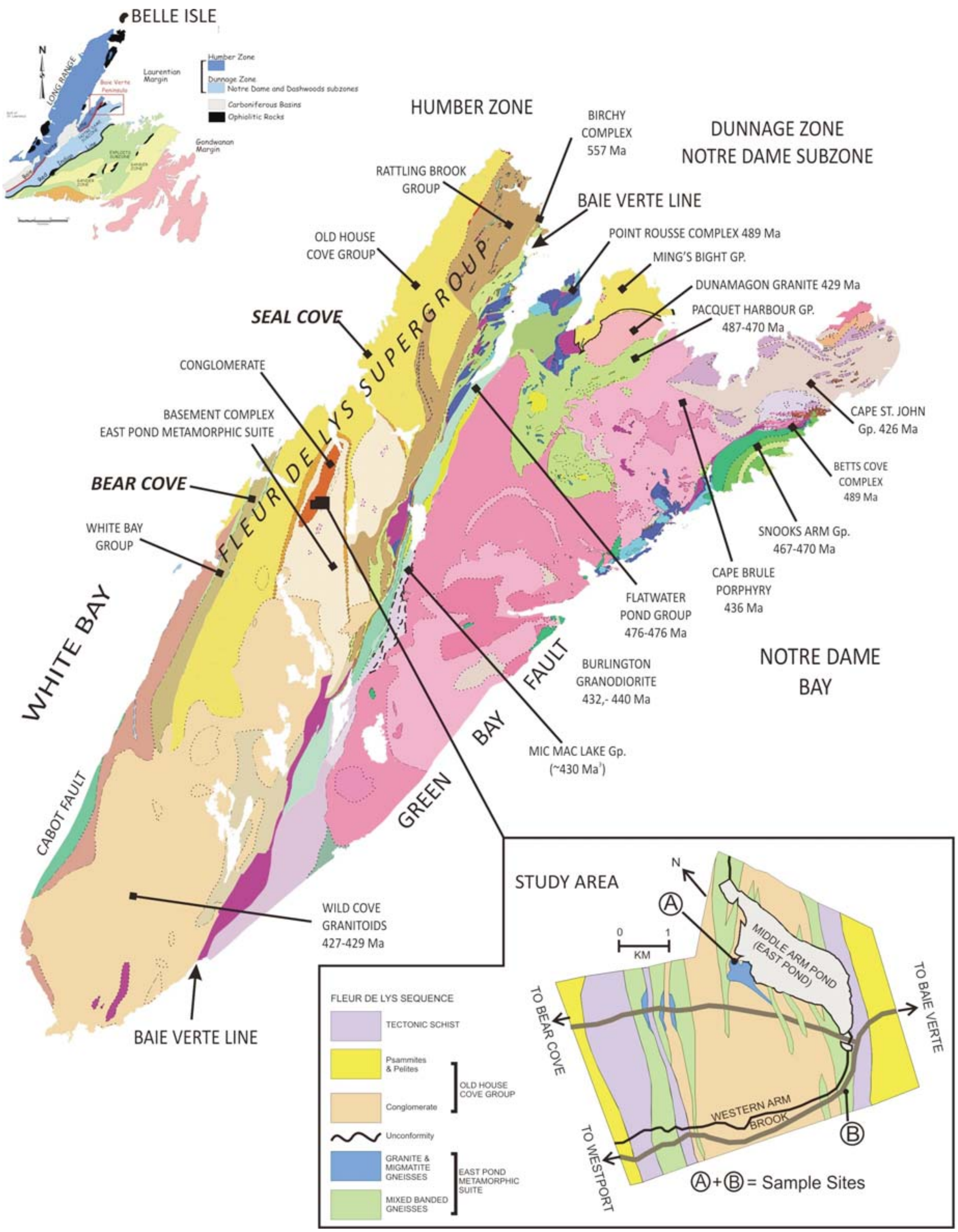


Figure 1. Simplified geology map of the Baie Verte Peninsula, NE Newfoundland, showing the distribution of the Fleur de Lys Supergroup (FdLS) to the west of the Baie Verte Line (suture). This suture separates the edge of the Laurentian Paleozoic continental margin (Humber Zone) in the west, from Paleozoic oceanic and arc terranes in the east (Notre Dame Subzone of the Dunnage Zone). Upper inset shows the broad tectonic domains of the Newfoundland Appalachians, with the Grenvillian basement of the Long Range (dark blue) overlain by allochthonous ophiolites (black). Note the position of Belle Isle some 20 km directly north of the Long Range. Lower inset shows the expanded study area where conglomerate and siliciclastic rocks of the Old House Cove Group unconformably overlie tectonic or antiformal wedges of basement gneiss (East Pond Metamorphic Suite); also shown is sub-vertical tectonic schist that separates the FdLS from the basement gneiss in many cases. U–Pb analyses on zircon from basement samples collected at sites A and B are described in this contribution. All other dates shown are from other sources describe in the text. Map modified from Skulski et al. 2010 and van Staal and Barr 2012.

mation of early Paleozoic ophiolites (cf. Dilek and Furnes 2011): the Lushs Bight Oceanic Tract (LBOT; with ophiolites dated between 510–501 Ma) and the Baie Verte Oceanic Tract (BVOT) comprising oceanic and island arc terranes, dated around 490 Ma, with overlying volcano-sedimentary sequences (480–467 Ma); and outboard farther still by a ribbon-like zone of continental crust of Laurentian affinity known as the Dashwoods microcontinent (Waldron and van Staal 2001; van Staal et al. 2009a, 2013; van Staal and Barr 2012). This composite terrane formed the basement to the 490–433 Ma Notre Dame arc (Fig. 1), collectively referred to as the Dunnage Zone. The FdLS is tectonically separated from the Dunnage Zone across the Baie Verte Line (BVL) (Fig. 1), a deformed, now subvertical, early Paleozoic suture. The precise tectonic evolution of the BVL and its relationships with the oceanic–arc assemblages of the BVOT to the east, and the FdLS to the west, including the late Neoproterozoic part of the Birchy Complex (ca. 556–563 Ma), is still poorly resolved (e.g. Kidd 1974; Bursnall 1975; Bursnall and de Wit 1975; Williams 1979; Hibbard 1983; Skulski et al. 2010; van Staal et al. 2013).

Between ca. 500 and 470 Ma, the passive Humber margin was progressively converted into a convergent margin. The Humber Zone and the westernmost part of the Dunnage Zone (the Notre Dame subzone, Fig. 1), which were deformed during the early Paleozoic Taconic orogeny, displays three distinct tectonic events between the Late Cambrian and Late Ordovician (495–450 Ma). Toward the end of this episode, the Notre Dame and Humber zones were intruded by several large Late Ordovician to Early Silurian granitoid plutons ranging in age between ca. 446–427 (e.g. Skulski et al. 2010; Fig. 1).

To summarize, the FdLS represents a critical sequence of transition between old continental crust to the west, and Paleozoic oceanic domains to the east. Understanding the framework and evolution of the basement to the FdLS is thus of importance for accurate restoration of the Paleozoic continental margin (e.g. the Humber Zone) and the Paleozoic ocean domains (e.g. the Dunnage Zone).

Geologic Framework of the FdL

A prominent metaconglomerate within the central part of the Fleur de Lys succession was discovered by Neale and Nash (1963), and later remapped in detail by de Wit (1972, 1974). Harland (1969) suggested that the conglomerate was probably a Varangian tillite. de Wit (1972) suggested the conglomerate was of fluvial origin and unconformably overlies a remobilized basement of gneiss and granite (de Wit 1980). These rocks were later renamed the East Pond Metamorphic Suite (EPMS) by Hibbard (1983) and incorporated by him (and Piasecki 1987, 1988) into the FdLS (i.e. the EPMS of Hibbard is the infrastructure of both the highly tectonized and metamorphosed basement and cover of the FdLS; Piasecki was less precise about the tectonostratigraphy, but incorporated the conglomerate into the EPMS basement, and separated it from the FdLS). The new U–Pb zircon dates described in this paper are from the central EPMS.

By contrast, de Wit (1972, 1980) mapped older basement sequences to the FdL region in two distinct tectonic environments: (1) a large elongate complex in the centre of the FdL sequence, and (2) as thin tectonic slivers representing modified tectonic units along White Bay. This basement was interpreted to underlie unconformably the Late Precambrian to Lower Ordovician siliciclastic rocks of the Fleur de Lys Supergroup (the cover). The cover consists of the above-mentioned basal conglomerate (including large boulders that display complex pre-entrainment tectonometamorphic fabrics; Fig. 2) overlain by a thick monotonous sequence of coarse psammite and semi-pelite intruded by large volumes of mafic dykes and sills, now mostly transformed into boudins of amphibolite and eclogite (de Wit and Strong 1975; Hibbard 1983) and lesser volumes of tholeiitic metabasalt. Collectively these coarse siliciclastic rocks and interlayered basalt sequences were known as the Seal Cove Group (later renamed the Old House Cove Group by Hibbard 1983; Fig. 1).

The Old House Cove Group is in turn conformably overlain by a relatively thin and varied fine to medium-grained siliciclastic–carbonate–volcanic succession of the Bear Cove Group (now known as the White Bay Group, Hibbard 1983) and Rattling Brook Group, respectively flanking White Bay in the west, and the Baie Verte Line in the east (de Wit 1972, 1974; Bursnall and de Wit 1975; Hibbard 1983; Skulski et al. 2010 van Staal et al. 2013; Fig. 1).

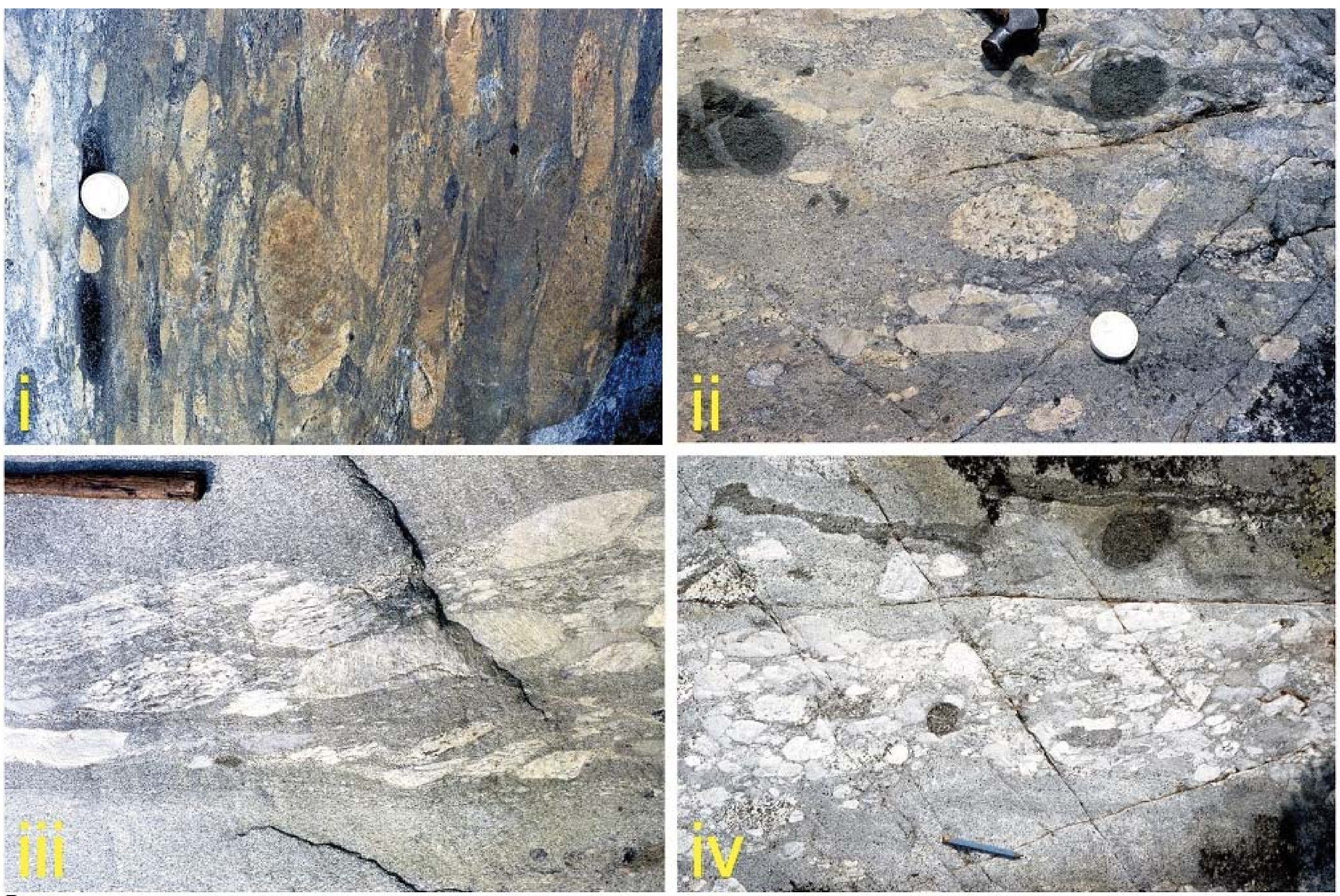
The FdLS was first subjected to intense deformation under conditions within the upper greenschist to amphibolite facies of metamorphism during the lower Paleozoic, likely in the Middle Ordovician (Bursnall and de Wit 1975; de Wit 1980; Hibbard 1983; Jamieson 1990; Cawood and Nemchin 2001; van Staal et al. 2009a, 2013; Skulski et al. 2010, van Staal and Barr 2012), although Church (1969) had previously suggested a pre-Early Ordovician age for this tectonism.

The regional structure of the FdL broadly conforms to the original suggestions of Neale and Nash (1963): a late (D_2) upright anticlinorium cored by a large elongate, northward-plunging basement massif, flanked to the east, north, and west by the Fleur de Lys metasedimentary rocks. The metasedimentary rocks overlie the basement massif and, using sedimentary structures, can be shown to young away from the basement (de Wit 1972, 1974, 1980). The contact between the basement and the FdLS is preserved in a series of (D_2) synformal wedges separated by wider antiforms in the centre domain of the basement, while the subvertical edges of the basement anticlinorium are marked by subvertical tectonic schist comprising a composite (D_1/D_2), predominantly flattening fabric (de Wit 1972, 1980; Piasecki 1987; Fig. 1). The east section of the FdLS (Rattling Brook Group and Birchy Complex) is dissected by east- and west-verging thrust zones (Kennedy 1975; Bursnall 1975; Hibbard 1983; Piasecki 1988; van Staal et al. 2013). Similar early tectonic zones with composite (D_1/D_2) fabrics are present in the White Bay Group (de Wit 1972; Hibbard 1983).

Both the basement and its cover were affected by at least two episodes of intense deformation, manifested throughout by polyphase

Figure 2. A: Large boulders of moderately deformed ortho- and para-gneisses (lower and upper boulders, respectively), and several granite boulders (top right), from the lowermost Fleur de Lys sequence that unconformably overlies older basement gneiss dated in this contribution. Note the pre-depositional internal gneissic fabrics (lower boulder) and folded/faulted veining (upper boulder) in the two large central boulders, illustrating their derivation from sources with complex, polyphase deformation and metamorphic histories prior to their entrainment in the Fleur de Lys Supergroup. Location: southern shore of lower Middle Arm. **B**: (i) Moderately deformed conglomerate sequence with rounded, clast-supported boulders, many with well-preserved pre-depositional gneissic fabric. (ii) Undeformed matrix-supported conglomerate sequence, with well-rounded granite boulder (centre) and two well-rounded amphibolite boulders (dark, near top). (iii) Slightly deformed layer of imbricated granitic and granitic gneiss boulders embedded in psammite; the lower psammite directly below the conglomerate displays fine mm thick layering (sub-parallel to hammer handle). (iv) Undeformed clast-supported conglomerate horizon with rounded to angular granitic, gneissic, psammite boulders and two well-rounded amphibolite clasts (dark). The conglomerate layer crosscuts faint bedding in the lower psammite (bedding; sub-parallel to the pencil), indicating younging towards the top of the photo.

A.



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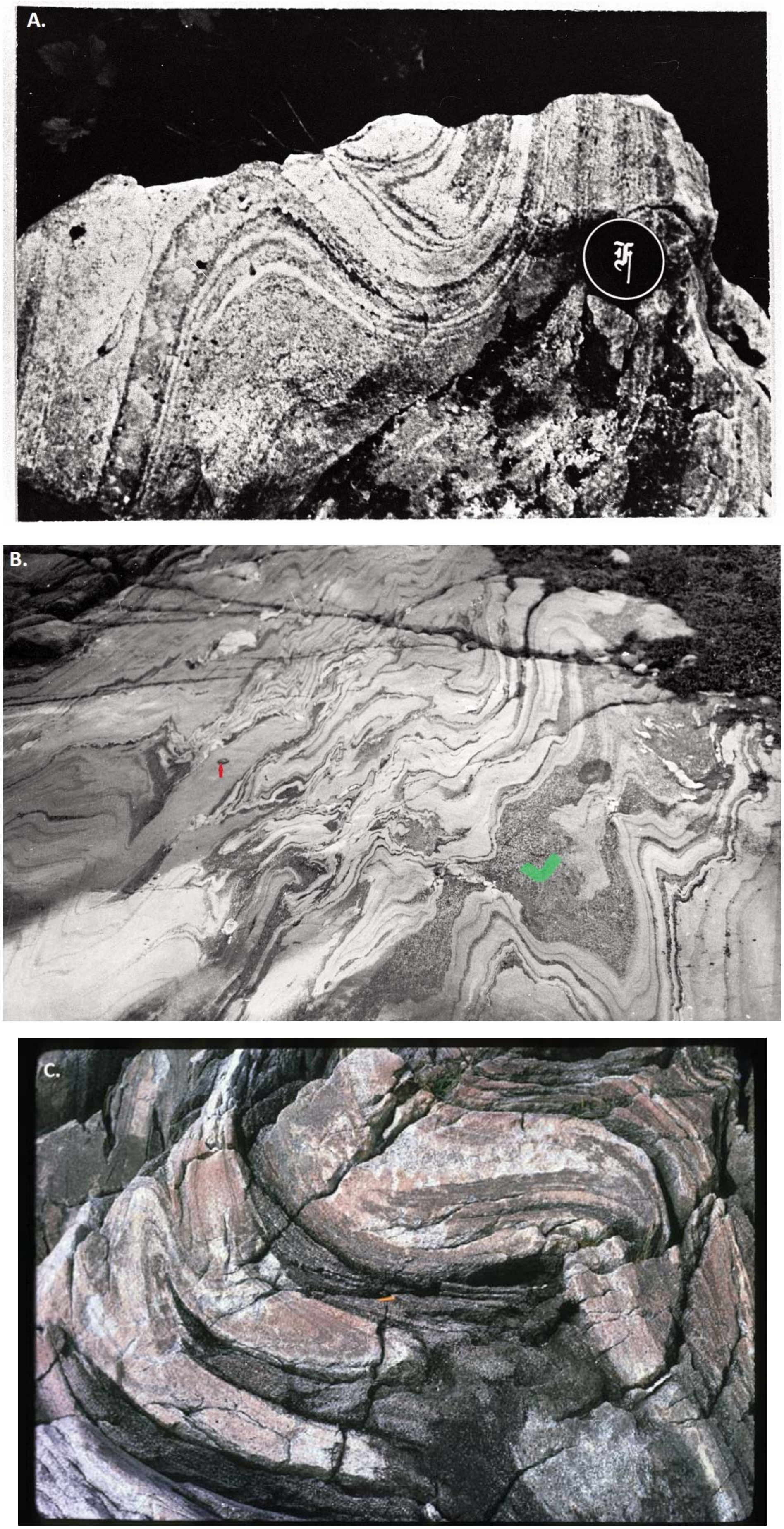

Figure 3. Comparative styles of polyphase coaxial folding in basement and Fleur de Lys cover rocks. **A**: early isoclinal folding of layering and foliation in basement gneiss (e.g. F_1/D_1, as seen in Figure 5A) refolded by upright open-tight D_2 folds with ill-defined axial planar cleavage (Westport Road close to location B on Fig 1). **B**: refolded bedding of Old House Cove Group pelite and psammite, with thin amphibolite layer (green tick) (red arrow points to camera lens cap for scale). Sub-vertical S_2 is axial planar to the second phase of open folding; S_1 is axial planar to the earlier isoclinal folds. Location: White Bay shore, between Eastern Arm and Middle Arm. **C:** Refolded D_1 isoclinal folds of interbedded feldspathic psammite (pale) and pelite (dark). Pelite displays strong axial planar S_1. Yellow pencil stub (center) for scale. Location: White Bay, at Middle Arm Rock (see detailed maps in de Wit 1972).

folding and the presence of a well-developed, near vertical schistosity (S_1), overprinted by a second, sub-vertical crenulation fabric (S_2). While the styles of folding in the basement and the cover were different, and it is not known with certainty if the ages of the episodes of folding in the basement and cover sequences are the same, the geometry and orientation of the polyphase structures are consistently similar in both sequences (Fig. 3).

The most comprehensive analyses of FdLS metamorphic conditions were carried out by Jamieson (1990). Her work on mineral assemblages from the central part of the FdLS record at least two major metamorphic events. During the first, early pressure and temperature conditions reached 7–8.5 kbar at about 450°C. These were subsequently overprinted by metamorphic conditions during ubiquitous feldspar and garnet porphyroblast growth at a lower pressure of ca. 6.5 kbar and higher temperatures up to 550–600°C. Thereafter substantial retrogression followed.

By contrast, the basement rocks (EPMS) also experienced early metamorphism, but at much higher pressures than the sedimentary FdLS cover, reaching pressures of at least 10–12 kbar at relatively low temperatures (different estimates suggest that the eclogite formed under a range of temperatures between 350–500°C), subsequently followed by re-equilibration at higher temperatures but lower pressures (700–750°C and 7–9 kbar).

Thus, the basement and cover rocks of the Humber zone on the Baie Verte Peninsula, west of the Baie Verte Line, experienced decompression (uplift and exhumation) from at least 33 km and 23 km to 23 km and 20 km, respectively, under an increasing geothermal gradient, and during which they were apparently brought into closer contact, implying that significant material was excised and that the central FdL terrane contains as yet poorly understood tectonic detachments, as implied by de Wit (1972, 1980), Hibbard (1983), and Piasecki (1988).

All the above-mentioned sequences were intruded during late- to post-deformation time, by the Wild Cove granitoid rocks, dated between 427 and 429 Ma (Fig. 1), part of a composite batholith now known as the Wild Cove Pond Igneous Suite (Hibbard 1983). The FdLS and its ductile deformation and multiple metamorphic history are thus older than mid-Silurian.

Basement Rocks to the FdL

The basement was remobilized, recrystallized, and in places structurally reconstituted during the early Paleozoic, and its original characteristics have frequently been obscured by the overprinted Paleozoic tectono-metamorphic events (de Wit 1980). The dominant remnant fabric in the basement is a gneissic layering, and at the vertical margins of the central basement complex, the overprinting is so intense as to cause complete reconstitution to tectonic schist (Fig. 1). The basement rocks were divided into (1) pre- and syn-layering rock types, and (2) post-layering rock types (de Wit 1972, 1980). The first group make up the fundamental architecture of the complex and consist mainly of supracrustal lithologies comprising about 70% of the basement gneiss. It is dominated by layered grey gneiss. Grey gneiss is interspersed with a sequence of mixed migmatitic-like banded gneiss that merges with pink granitic gneiss associated with migmatite. Clear contacts are rarely observed. One area was found to contain well-preserved pink granitic gneiss. The best exposed examples occur as a small discrete body along the west shore of Middle Arm Pond (East Pond of Hibbard 1983; Fig. 1, lower inset). It has a zone of migmatite towards the southwest, and includes remnants of the banded gneiss. Similar migmatite also occurs around the margins of Pine Pond.

de Wit (1972, 1980) interpreted the field relationships to indicate that the granitic gneiss intruded the basement gneiss, whereas Hibbard (1983) described the granitic rocks to have intruded the rocks of the Old House Cove Group. By contrast, Piasecki (1987, 1988) believed the lower FdLS rocks, including the conglomerate, are migmatized and, because he did not encounter clasts with tectonic fabrics earlier than those in the matrix of the conglomerate units, did not accept the presence of an older basement to the FdLS. Piasecki (1987) specifically stated most of the metaconglomerate is migmatized. Figure 2 shows a large number of clasts of conglomerate with pre-depositional fabrics at high angles to the D_1/D_2 tectonic fabrics in the matric and schist zones (for more examples, see also de Wit 1972); the conglomerate layers show no signs of migmatization. Notable too is the presence of metabasalt (amphibolite) clasts, likely reworked from basaltic lavas in the lower FdLS. The interested reader should consult de Wit (1972, 1974) and Hibbard (1983) for detailed descriptions of their field observations; the variations of deformation of the conglomerates from intense to subtle; and the petrography of these rocks. Below we provide a short synopsis before we test these relationships with geochronology.

Paragneiss

The best exposures of grey paragneiss appear east of the junction of the Bear Cove and Westport Roads, forming the first 3 km of basement exposures along the Westport Road. These comprise the greater portion of the gneiss. A prominent subvertical layering is the most striking structure (Fig. 4). One of our samples analysed here is derived from there (sample site B, Figs. 1 and 4).

In the field these rocks vary from massive, fine grained, greenish-grey rocks to well-foliated flaggy types. The massive types are subordinate. Layering and a sub-parallel foliation are extremely regularly developed, usually on a 1 mm to 10 cm scale. Some massive layers are up to 50 cm thick. The layering is delineated by variable concentrations of muscovite or biotite and epidote, separating usually thicker, quartz-enriched bands. S_1 is outlined by small oriented biotite and muscovite flakes, commonly slightly oblique to the layering. Plagioclase is subordinate, never exceeding 20% by volume, and usually clouded. Microcline may be present in amounts up to 5–10%, in places as augen. Rare quartzite together with pelite, epidosite and amphibolite occur as subordinate layers.

Gneiss may be seen interlayered with very regular 1–10 cm-layered leucocratic rocks of granitic to granodioritic composition. Individual layers are traceable for tens of metres. The leucocratic layers separate grey gneiss on a regular and similar thickness variation. Contacts are sharp. The thinner (millimetre) leucocratic bands are discontinuous, but stretch over several metres. In places the layering is isoclinally folded, (Fig. 5A), sometimes displaying an axial planar cleavage, which in turn is refolded (rarely seen, e.g. Fig 3A) by folds that have a composite axial planar cleavage (S_1/S_2), inferred to be the composite regional Paleozoic fabric. The regular layering is in contrast to mixed, migmatitic layered gneiss near pink granitic gneiss around Middle Arm Pond (East Pond) and on the Bear Cove Road (Fig. 1).

Pink Granite Gneiss

Mixed migmatitic-like layered gneiss gradationally increases in leucocratic pink layers towards the pink granite gneiss, until only dark melanocratic folia, blobs, and wisps (down to 1 mm thick) remain. In a few small domains (< 1 m) the layering is completely lost. This is best seen in the discrete pink granite gneiss body along Middle Arm

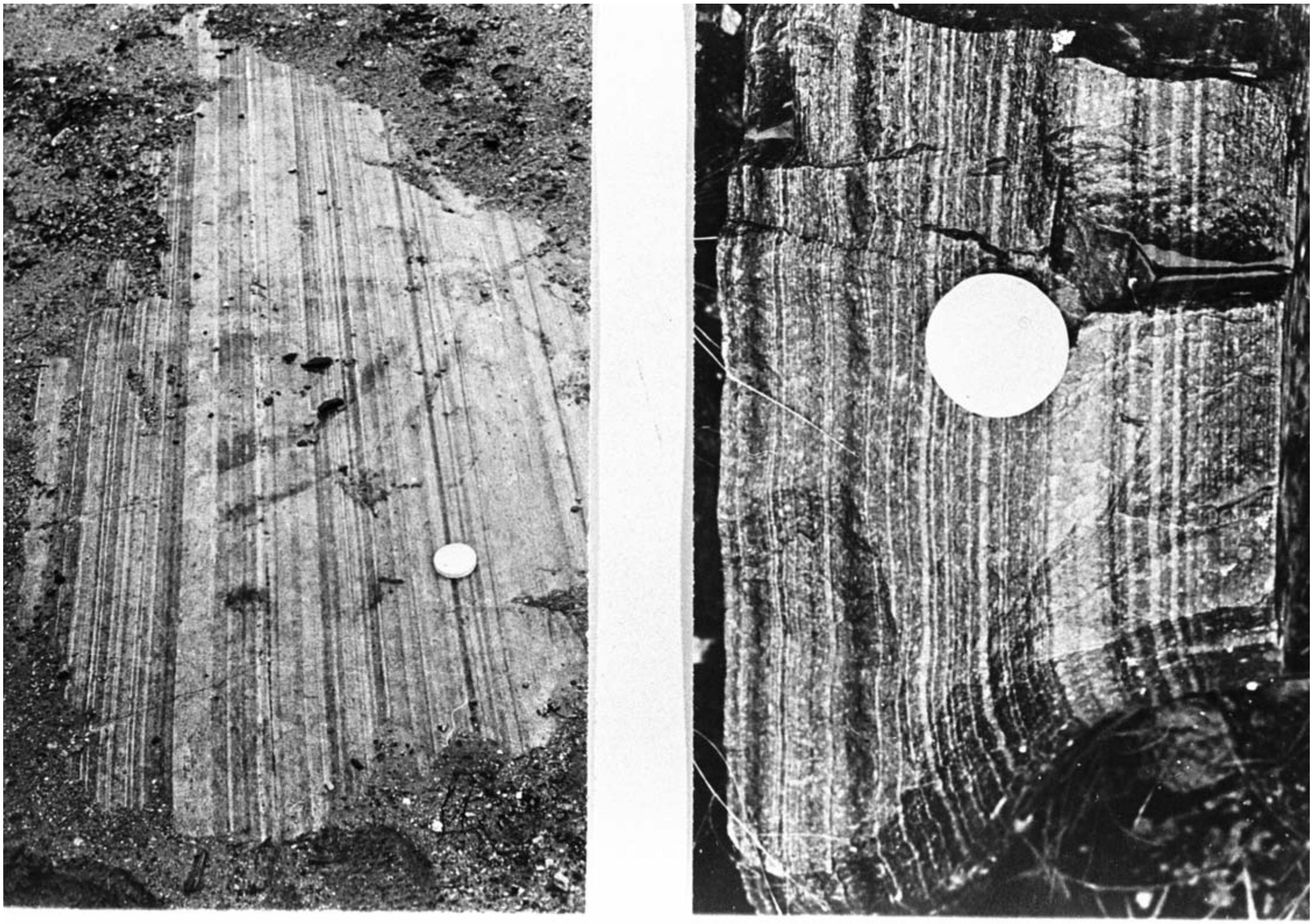

Figure 4. Left, outcrop of banded white granite-granodiorite composition interlayered with darker grey gneiss. Sample on the right (01-1340) was selected for U–Pb zircon analyses. Location: Locality B, Figure 1.

Pond. Due to the inhomogeneous nature of the deformation in this area, this body has been remarkably well protected from Paleozoic deformation (de Wit 1974, 1980). Here, the granite contains 20–40% microcline of the total 65–75% feldspar content. A sample from this rock, at locality A (Fig. 1) is the second of our dated rocks.

Aplite
Quartzo-feldspathic aplite dykes, veins, and pegmatites are seen at several localities. The best examples are along the Bear Cove Road, cutting mixed layered gneiss, for example at Locality B (Fig. 1, lower inset) about 6 km from the Baie Verte Road. The dykes are clearly intrusive, cutting the layering at a shallow angle (Fig. 5A). Margins are sharp against gneiss, and there is no evidence of marginal chilling. The dykes do not display any internal layering, but they are crosscut by S_1, manifested by the alignment of small biotite, muscovite and chlorite laths. A xenolith of mixed gneiss in one of these dykes confirms that the formation of the leucocratic/melanocratic gneissic layering predates S_1 (Fig. 5B and see, respectively, figure 7 and Plate 4-6 in de Wit 1980; and Hibbard 1983).

Small pink ptygmatic veins and pegmatites occur throughout the basement, but are especially well developed in the grey gneiss and mixed layered gneiss. They cut the layering, but are frequently folded with S_1 axial planar but do not occur in the surrounding Fleur de Lys metasedimentary rocks. Both the aplite dykes and the pegmatites are thought to represent the final stages of pre-Fleur de Lys activity in the basement complex, but their absolute age has not been determined.

Eclogite and Amphibolite
Eclogite and amphibolite occur in many boudins and lenses throughout the main basement massif. The boudins are never larger than 10 m in diameter and are usually much smaller. They represent deformed and metamorphosed mafic dykes that cut across the basement–Old House Cove Group interface and are thus of late Precambrian to earliest Paleozoic age. Where the dykes cut across the basement rocks they frequently preserve eclogite mineralogy. A retrograded eclogite has been dated at 464 ± 13 Ma by U–Pb on zircon using SHRIMP (van Staal et al. 2009b).

The metamorphic conditions under which the eclogite formed, under relatively anhydrous conditions, range between 8 and 12 kbar pressure, and 350–450°C temperature (e.g. de Wit and Strong 1975), similar to the

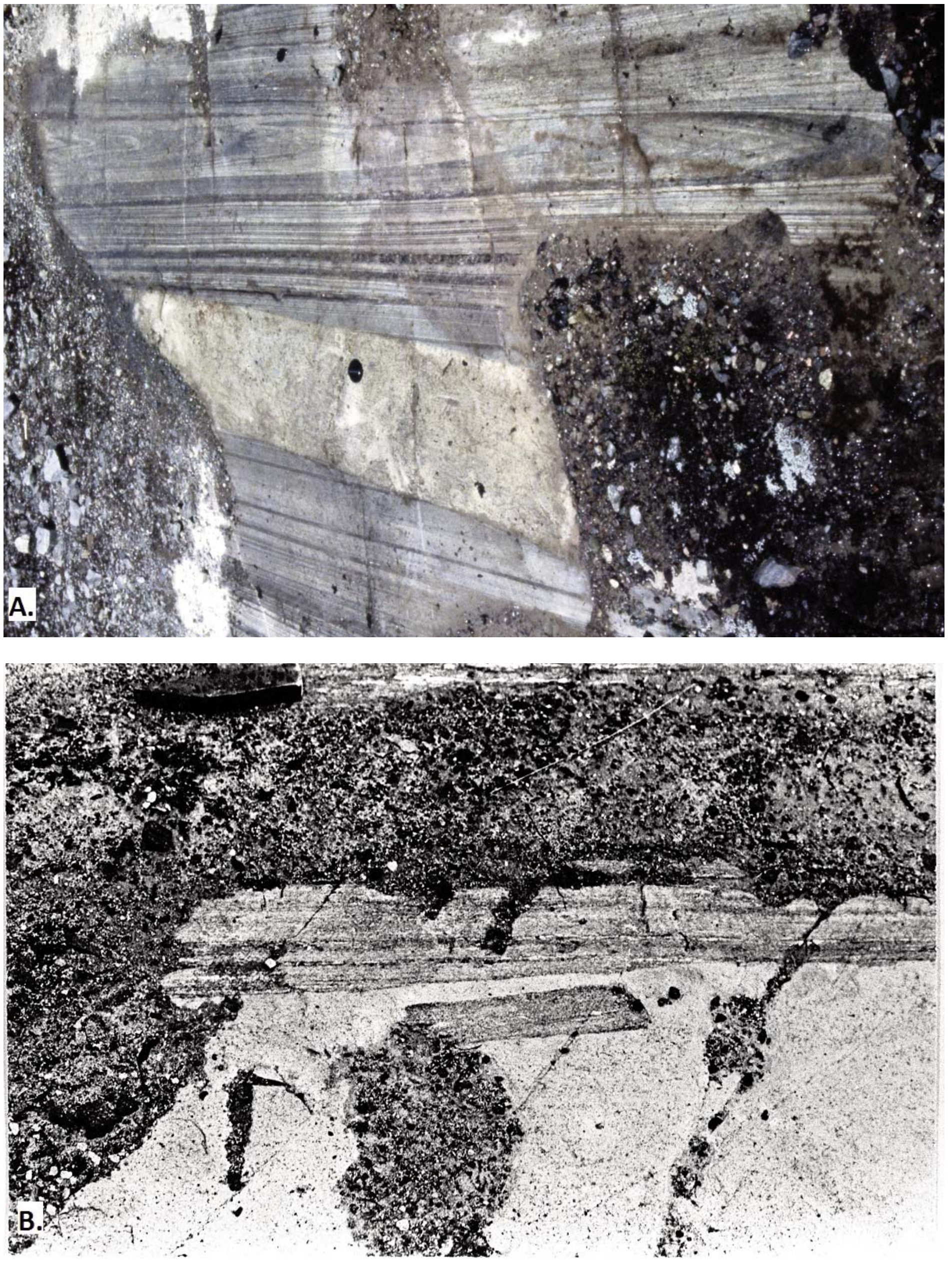

Figure 5. **A**: mixed layered gneiss with isoclinal D_1 fold in the pre-Fleur de Lys basement. Layering is cut by a late, but pre-D_2, felsic dyke (white) that crosscuts both limbs of the isoclinal folded layering (not seen on the photo). Location is close to Figure 4. **B**: Xenolith of mixed layered gneiss in felsic dyke of A.

general basement *P-T* conditions documented by Jamieson (1990). The preservation of the low-temperature eclogite suggests that metamorphism involved relatively cold continental crust and thus predated tectonic reworking during subsequent higher temperature events (Jamieson 1990). Their field relations, petrography, petrology, and geochemistry have been described elsewhere (de Wit and Strong 1975; Jamieson 1990).

GEOCHRONOLOGY OF THE FLEUR DE LYS FLANKING THE BAIE VERTE LINEAMENT

U–Pb analyses on detrital zircon from an upper FdLS psammite unit adjacent to the Birchy Complex (Flat Point Formation) has yielded mostly Mesoproterozoic Grenvillian ages (1.0–1.5 Ga), typical of Laurentian basement ages in the Grenville province of southern Labrador. A few grains have yielded Paleoproterozoic (1.8–2.0 Ga) and Neoarchean (2.4–3.0 Ga) dates. The youngest detrital zircon has a concordant U–Pb date of 990 ± 52 Ma, implying a maximum age of early Neoproterozoic sedimentation along the passive margin of the Humber Zone. However, the detrital zircon grains in samples of the Kidney Pond conglomerate just to the east of the BVL, yielded U–Pb zircon ages ranging from ca. 2550 to 550 Ma, and associated felsic rocks have inherited zircon as old as 2600 Ma (Skulski et al. 2010). This suggests a maximum age of late Neoproterozoic clastic sedimentation along the passive margin of the Humber Zone.

General geologic evidence suggests that the main phase of metamorphism in the western Baie Verte Peninsula predated 420 Ma, and a U–Pb zircon date of ca. 434 Ma from a pegmatite affected by an early (S_1) fabric places a younger limit on the early FdLS deformation in Corner Brook to the south of the Burlington Peninsula (see Cawood et al. 1994; Waldron et al. 1998; Cawood and Nemchin 2001). However, elsewhere, early deformation in the Humber Zone has been estimated to have occurred between 477 and 459 Ma, based on $^{40}Ar/^{39}Ar$ analyses (S. Castonguay et al. unpublished data 2009; reported in Skulski et al. 2010). Since the argon data likely record cooling ages, this suggests a minimum age of about 460 Ma for the early deformation and high pressure metamorphism, consistent with the metamorphic age of eclogite (see below).

The early deformation is interpreted to be related to obduction of ophiolite complexes and arc collision of the BVOT and the FdLS terrain during the first phase of the early Ordovician Taconic Orogeny (e.g. Skulski et al. 2010; van Staal et al. 2009a, 2013). Since this tectonic emplacement induced subsidence of the foreland basin of the Humber Zone to the east of the FdL, the foreland basin stratigraphy provides age information on tectonic loading of the former continental margin. Rapid subsidence sedimentation began in the Dapingian (ca. 470 Ma – see Waldron et al. 1998 [Goose Tickle Group]), was interrupted during the Sandbian, and resumed rapidly thereafter to continue into the Llandovery (ca. 458–430 Ma), which also likely records a minimum age for the early high-pressure metamorphism at the leading edge of the former continental margin that occurred at depths of ca. 30–40 km (Jamieson 1990; Skulski et al. 2010; van Staal et al. 2013).

A metamorphic zircon inclusion in the rim of garnet of a retrograded eclogite from the FdL basement at Gull Pond has been dated at 464 ± 13 Ma (van Staal et al. 2009b). The zircon probably grew during retrogression to amphibolite because its REE chemistry suggests it was in equilibrium with hornblende rather than the eclogite assemblage (C. van Staal, pers. comm. 2013, 2014). If correct, this date provides a possible age for the second metamorphic phase, and thus also a minimum age for the earlier eclogite metamorphism and deep tectonic burial of the basement and its lower FdLS cover under a relatively low geothermal gradient.

The high temperature metamorphic peak was followed by either rapid cooling and exhumation of the upper Fleur de Lys Supergroup by ca. 465 Ma ($^{40}Ar/^{39}Ar$ dates on muscovite and hornblende in the Birchy Complex; van Staal et al. 2013) and/or relatively slow cooling of the lower sequences before 429–417 Ma, as suggested by $^{40}Ar/^{39}Ar$ dates on muscovite and hornblende from the EPMS (reported in Jamieson 1990).

Geochronology of Two FdL Basement Samples

Two basement samples, one from the grey layered paragneiss, and one from the pink granite were selected for U–Pb zircon analyses by SHRIMP (Fig 1.)

Analytical Techniques

Zircon grains were separated at the Research School of Earth Sciences (RSES) of the Australian National University (ANU), Canberra, using conventional density and magnetic techniques. The zircon concentrates were hand-picked under a binocular microscope and mounted in epoxy together with grains and chips of the RSES reference zircon samples FC1 and SL13. Mounted grains were polished to approximately half their thicknesses to expose internal structures. In preparation for the SHRIMP analyses, the sectioned zircon grains were then photographed in transmitted and reflected light pairs which, together with SEM cathodoluminescence (CL) images, were used to decipher the internal structures of the sectioned grains and to target specific areas within the zircon grains for spot analysis.

U–Th–Pb analyses were made using both SHRIMP II and SHRIMP RG at the RSES. In order to obtain an accurate U–Pb calibration for each analytical session, analyses of the standard FC1 were interspersed with those of the unknowns throughout the sessions, usually with 1 standard analysed for every three unknowns. Each analysis consisted of 6 scans through the mass range $^{196}Zr_2O$ (2 seconds count time), ^{204}Pb (10 seconds), mass 204.1 (background, 10 seconds), ^{206}Pb (10 seconds), ^{207}Pb (30 seconds), ^{208}Pb (5 seconds), ^{238}U (5 seconds), ^{248}ThO (5 seconds), and ^{254}UO (2 seconds). The data were reduced in a manner similar to that described by Williams (1998, and references therein), using the SQUID 1 Excel Macro of Ludwig (2000) or SQUID 2 (Ludwig 2009). For the zircon age calibration the Pb–U ratios were normalized relative to a value of 0.1859 for the $^{206}Pb/^{238}U$ ratio of the FC1 reference zircon samples, equivalent to an age of 1099 Ma (Paces and Miller 1993). Uncertainties in the Pb–U calibration from the various analytical sessions are included in computed $^{206}Pb/^{238}U$ ages or concordia ages (Ludwig 1998), but are not included in the individual analyses as reported in the data tables (errors are, however, given separately in the tables). Corrections for common Pb were made using the appropriate model values of Stacey and Kramers (1975). U and Th concentrations were determined relative to those measured in the SL13 standard (Claoué-Long et al. 1995).

Uncertainties reported for individual analyses (ratios and ages) are at the 1σ level, but the uncertainties in calculated weighted mean ages or upper intercept ages are reported as 95% confidence limits. Concordia plots, regressions and any weighted mean age calculations were carried out using Isoplot 3.00 (Ludwig 2003).

Results

Sample 01-134 (Locality B): Paragneiss

The zircon samples extracted from this paragneiss (Fig. 4) show a large range in sizes from ca. 500 μm (in the longest dimension) to small 50 μm grains. All show some degree of rounding, both from mechanical abrasion/fracturing and metamorphic rounding. The internal structures are complex with cathodoluminescence imaging (CL) showing reworked older cores rimmed or embayed by later phases (Fig. 6). Metamorphic zircon is present as rims or as discrete, rounded 'soccerball' grains commonly observed in high-grade metamorphic populations.

As expected from the CL imaging, the SHRIMP U–Pb data are complex and difficult to interpret in terms of establishing a maximum age of deposition for this paragneiss and the timing of metamorphism (Table 1). Cores and rounded grains plot as concordant data between ca. 2580 and 2820 Ma, with the vast majority plotting in a more restricted age range between 2650 and 2750 Ma (Fig. 7 A-D). The discordant data appear to mainly fall along an intermediate Proterozoic Pb-loss trend, probably complicated by more than one episode of Pb loss. The next group of concordant data are clustered on and below concordia between ca. 1750 and 1850 Ma. These spots were exclusively sited in bright-CL overgrowths or discrete grains predominantly showing the typical 'soccerball' shape associated with high-grade metamorphic zircon. Zoning patterns are non-existent, very faint or reworked. Although these zircon grains do not all have the characteristic low Th/U geochemistry common seen in metamorphic zircon, this could simply be due to the absence of a Th-bearing crystallizing phase during metamorphism. The youngest group of data plot close to concordia and appear to define a Mesoproterozoic (Grenville) age around 1000–1100 Ma. It is difficult to calculate an upper intercept date, as choosing which data to include in such a calculation is subjective given the discordant nature of the data, but one model suggests an upper intercept around 1300 Ma (Fig. 7C). No zircon younger than ca. 1000

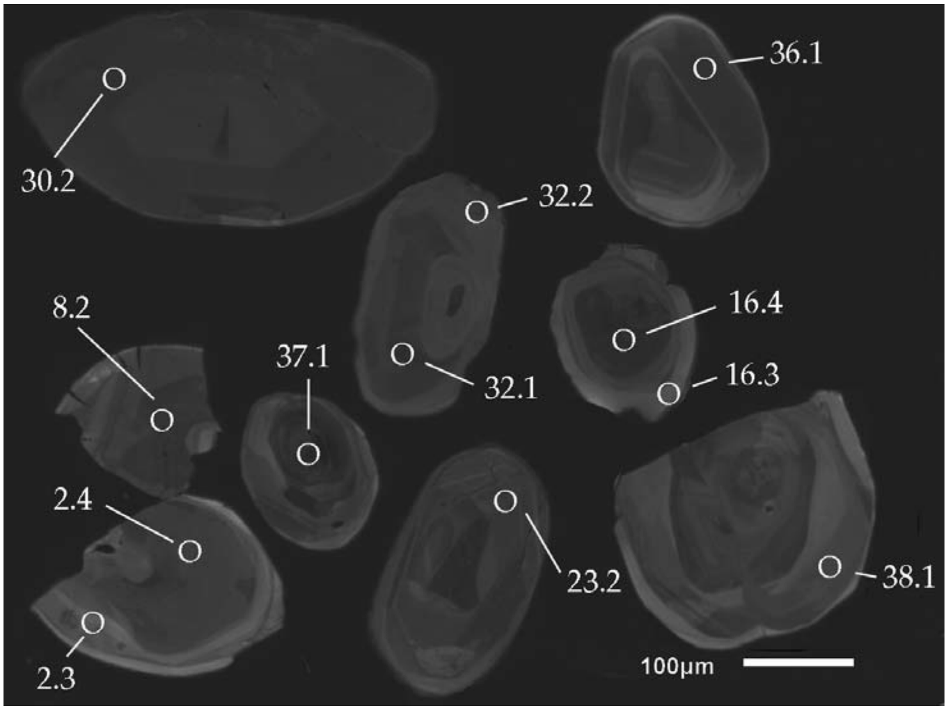

Figure 6. Cathodoluminescence image (CL) of rounded zircon grains from sample 01-134. Internal structures are complex with reworked older cores rimmed or embayed by later zircon phases. Metamorphic zircon is present as rims or as discrete, rounded 'soccerball' grains. See text for further information.

Ma was recorded. Metamorphic growth clearly identifiable through CL is impossible to date due to the extremely low U contents (<0.01 ppm).

Although the above results are consistent with a combined detrital source from the Superior Province and a Grenvillian sequence, the data do not prove this paragneiss could not have been deposited at a younger time; our limited data may merely reflect non-preservation of younger detrital zircon in this small sample.

Sample 01-133 (Locality A, Fig. 1): Pink Granite

Most zircon grains from this sample are pale pink, clear and tend to have smooth, rounded edges and tips consistent with metamorphic reworking and growth. Cathodoluminescence images show remnants of zoned magmatic zircon as cores surrounded by both dark- and light-CL zircon rims (Fig. 8A). It is possible that these represent two phases of post-igneous growth, with the inner (older) areas generally having relatively dark CL response and limited or no zoning. The outer rims are narrow, relatively light-CL and without internal structure.

SHRIMP spot selection was aimed at dating all observed phases of growth. The concordia plot of the zircon U–Pb data for this sample shows three clusters, with the remainder scattering between these groups (Table 2; Fig. 8B).

The oldest cluster is best defined by an upper intercept date of 1491 ± 19 Ma; calculated from the 6 more concordant points only. The zircon grains from this group comprise magmatic zoned grains or cores mantled and embayed by metamorphic overgrowths.

The second cluster of data defines an intermediate age of 966 ± 33 Ma (calculated as a weighted mean $^{207}Pb/^{206}Pb$ date on 6 points). Characterization of these zircon grains as igneous or metamorphic is difficult, but the general cross-cutting and structureless form suggests a metamorphic origin.

The third and youngest group comprises a cluster of five analyses located in small metamorphic overgrowths. These overgrowths have low U, variable but occasionally extremely low Th/U (0.002) and give a weighted mean $^{206}Pb/^{238}U$ date of 482.9 ± 8.5 Ma. This Tremadocian age clearly records a metamorphic event that predated the bulk of the Ar/Ar estimates of 468–459 Ma for D_1 deformation by at least 10–20 m.y., and the retrograde amphibolitization of eclogite by a similar age range. It is therefore a more likely age of the earliest tectonic burial and metamorphism in the FdL region (e.g. ca. 483 Ma).

CORRELATION OF FLEUR DE LYS ROCKS WITHIN THE HUMBER ZONE

Basement Correlative Units

Although the Long Range Inlier has long been recognized as having an affinity with the Grenville province of the nearby Laurentian Shield (e.g. Clifford and Baird 1962), the most comprehensive isotopic analysis of the Grenvillian basement exposed along the Long Range was by Heaman et al. (2002), and comparisons are made below with the results documented in that study. In summary, amphibolite- to granulite-facies granitic to granodioritic orthogneiss with screens of dismembered paragneiss occur throughout the Long Range. The remainder is underlain by Mesoproterozoic intrusions, the oldest of which are Pinwarian (ca. 1500 Ma), while the younger Grenvillian plutons range in age between 1100 and 1000 Ma. U–Pb results for both zircon overgrowths in basement gneiss units and titanite in older Grenvillian granitoid intrusions indicate a late high-grade metamorphism around 970 Ma.

Detrital zircon grains between 1000–1100 Ma from the FdL paragneiss are consistent with similar dates from an early Grenvillian granitoid suite (Group I) and the general Grenvillian metamorphism exposed on the Long Range, in particular that recorded by the Lake Michel composite intrusion (the largest plutonic unit in the Long Range Inlier, exposed over an area of more than 1500 km^2; Heaman et al. 2002). Lower intercept ages also overlap, although these are slightly older in the FdL sample (770 ± 210 Ma).

The date of the magmatic zircon from the granite sample of the FdL basement (1491 Ma) is similar to igneous zircon of many pre-Grenvil-

Table 1: Summary of SHRIMP U–Pb zircon data for metasediment sample 01-134 (LAB2)

Grain Spot	% $^{206}Pb_c$	ppm U	ppm Th	$^{232}Th/^{238}U$	(1) ppm $^{206}Pb^*$	(1) $^{206}Pb/^{238}U$ Age (Ma)	(1) $^{207}Pb/^{206}Pb$ Age (Ma)	% Dis-cor-dant	(1) $^{207}Pb^*/^{206}Pb^*$	±%	(1) $^{207}Pb^*/^{235}U$	±%	(1) $^{206}Pb^*/^{238}U$	±%	err corr
1.3	0.02	105	140	1.38	41.0	2419 ±23	2590 ±19	+8	0.1733	1.1	10.881	1.6	0.4554	1.2	0.72
1.4	--	66	69	1.09	28.9	2672 ±33	2690 ±8	+1	0.1841	0.5	13.038	1.6	0.5137	1.5	0.96
2.3	0.30	25	31	1.29	7.1	1839 ±25	2206 ±23	+19	0.1383	1.3	6.295	2.0	0.3301	1.6	0.76
2.4	0.05	87	49	0.58	16.3	1275 ±18	1654 ±23	+25	0.1016	1.3	3.063	2.0	0.2186	1.6	0.78
7.2	0.35	28	2	0.07	8.0	1841 ±45	1827 ±28	-1	0.1117	1.5	5.088	3.2	0.3305	2.8	0.88
8.2	--	91	67	0.76	12.4	948 ±18	1058 ±35	+11	0.0746	1.8	1.629	2.7	0.1584	2.0	0.75
9.2	0.04	46	63	1.41	16.2	2200 ±25	2373 ±25	+9	0.1524	1.5	8.549	2.0	0.4067	1.3	0.67
10.2	0.03	82	48	0.60	26	2006 ±21	2370 ±33	+18	0.1521	1.9	7.656	2.3	0.3651	1.2	0.53
11.2	0.07	67	117	1.81	26.1	2422 ±36	2537 ±10	+5	0.1679	0.6	10.561	1.9	0.4561	1.8	0.95
12.4	0.19	100	126	1.31	17.6	1199 ±23	1465 ±21	+20	0.0919	1.1	2.590	2.4	0.2044	2.1	0.88
13.2	0.05	55	98	1.86	23.5	2616 ±29	2712 ±22	+4	0.1865	1.3	12.870	1.9	0.5004	1.3	0.71
14.2	--	13	11	0.82	4	1795 ±43	1708 ±33	-6	0.1047	1.8	4.632	3.3	0.3210	2.7	0.84
15.2	0.32	129	147	1.18	24	1258 ±13	1638 ±26	+25	0.1007	1.4	2.994	1.8	0.2156	1.1	0.62
16.3	0.56	21	18	0.89	3.3	1098 ±19	1147 ±86	+5	0.0780	4.3	1.997	4.7	0.1857	1.9	0.40
16.4	0.12	143	120	0.87	21.5	1036 ±11	1125 ±21	+8	0.0771	1.1	1.855	1.6	0.1744	1.2	0.74
17.2	0.06	29	35	1.26	7.5	1712 ±24	1728 ±25	+1	0.1058	1.4	4.436	2.1	0.3042	1.6	0.76
18.2	0.12	33	47	1.48	11.2	2151 ±38	2379 ±17	+11	0.1529	1.0	8.352	2.3	0.3961	2.1	0.90
19.2	--	117	135	1.19	17	999 ±11	1073 ±19	+7	0.0752	0.9	1.738	1.5	0.1677	1.2	0.78
20.2	0.08	74	47	0.65	26	2236 ±43	2522 ±20	+13	0.1664	1.2	9.516	2.6	0.4147	2.3	0.88
22.2	0.01	119	80	0.69	51.1	2611 ±40	2660 ±16	+2	0.1807	1.0	12.447	2.1	0.4995	1.9	0.89
23.2	--	81	84	1.07	30.0	2308 ±31	2519 ±9	+10	0.1661	0.5	9.859	1.7	0.4305	1.6	0.95
24.3	0.03	30	56	1.91	14.4	2836 ±54	2768 ±21	-3	0.1930	1.3	14.707	2.7	0.5526	2.4	0.88
24.4	--	212	130	0.63	84.0	2446 ±23	2609 ±19	+7	0.1753	1.1	11.152	1.6	0.4614	1.1	0.70
25.3	0.08	86	52	0.62	39	2744 ±27	2737 ±7	-0	0.1894	0.4	13.857	1.3	0.5305	1.2	0.95
26.2	0.01	346	43	0.13	164	2836 ±25	2783 ±3	-2	0.1948	0.2	14.840	1.1	0.5526	1.1	0.98
27.2	0.10	61	94	1.60	25	2560 ±34	2538 ±20	-1	0.1681	1.2	11.299	2.0	0.4876	1.6	0.80
28.2	--	66	65	1.03	28	2597 ±43	2656 ±18	+3	0.1804	1.1	12.339	2.3	0.4962	2.0	0.88
29.2	0.06	24	4	0.18	6	1706 ±24	1705 ±25	-0	0.1045	1.3	4.364	2.1	0.3030	1.6	0.76
30.2	0.08	73	79	1.12	30.2	2533 ±27	2631 ±9	+5	0.1776	0.6	11.787	1.4	0.4813	1.3	0.91
31.1	0.11	63	64	1.04	14.9	1559 ±17	1732 ±17	+11	0.1060	0.9	4.001	1.6	0.2737	1.3	0.80
32.1	--	81	2	0.03	12.4	1060 ±18	1128 ±25	+7	0.0773	1.2	1.904	2.2	0.1787	1.8	0.83
32.2	0.08	34	1	0.03	6.1	1210 ±17	1270 ±36	+5	0.0830	1.8	2.365	2.4	0.2065	1.5	0.65
33.2	--	54	78	1.51	16.6	1983 ±34	2123 ±24	+8	0.1319	1.4	6.549	2.4	0.3601	2.0	0.82
35.1	0.14	65	124	1.96	24.6	2342 ±35	2486 ±12	+7	0.1629	0.7	9.838	1.9	0.4380	1.8	0.93
36.1	0.30	59	64	1.12	17.8	1930 ±22	2207 ±24	+15	0.1384	1.4	6.658	1.9	0.3490	1.3	0.70
37.1	0.06	150	107	0.73	68.0	2726 ±34	2701 ±14	-1	0.1853	0.9	13.454	1.8	0.5264	1.5	0.87
38.1	0.41	39	79	2.12	11.5	1928 ±45	2306 ±29	+19	0.1466	1.7	7.044	3.2	0.3485	2.7	0.85
39.1	0.35	31	9	0.31	9.1	1869 ±38	2001 ±28	+8	0.1230	1.6	5.706	2.8	0.3364	2.3	0.83
40.1	0.23	9	10	1.14	2.4	1708 ±38	1702 ±58	-0	0.1043	3.1	4.361	4.0	0.3033	2.5	0.63
41.1	0.00	48	10	0.22	12	1618 ±19	1706 ±17	+6	0.1045	0.9	4.111	1.6	0.2853	1.3	0.82
42.1	0.01	165	46	0.29	74	2710 ±40	2761 ±17	+2	0.1922	1.1	13.850	2.1	0.5225	1.8	0.86
43.1	0.10	63	49	0.81	24	2346 ±50	2521 ±40	+8	0.1664	2.4	10.069	3.5	0.4390	2.5	0.72
44.1	--	36	52	1.52	13	2341 ±40	2567 ±12	+10	0.1709	0.7	10.320	2.2	0.4379	2.0	0.95
45.1	0.26	32	12	0.40	9	1800 ±23	1707 ±25	-6	0.1046	1.4	4.643	2.0	0.3220	1.5	0.73
46.1	0.00	66	83	1.30	23	2158 ±23	2359 ±10	+10	0.1512	0.6	8.286	1.4	0.3976	1.2	0.91
47.1	--	39	65	1.70	13	2063 ±34	2264 ±17	+10	0.1430	1.0	7.436	2.2	0.3771	1.9	0.89
48.1	0.05	34	38	1.14	13	2413 ±29	2537 ±12	+6	0.1680	0.7	10.512	1.6	0.4539	1.4	0.89
49.1	0.00	35	62	1.81	15	2627 ±62	2679 ±26	+2	0.1829	1.6	12.687	3.3	0.5031	2.9	0.87
50.1	0.05	121	211	1.81	54	2706 ±25	2706 ±12	+0	0.1859	0.7	13.372	1.4	0.5217	1.1	0.85
51.1	0.03	193	74	0.40	77	2463 ±26	2619 ±8	+7	0.1763	0.5	11.314	1.4	0.4654	1.3	0.93
52.1	0.02	102	101	1.02	39	2371 ±48	2557 ±26	+9	0.1700	1.6	10.416	2.9	0.4445	2.4	0.84
53.1	0.01	100	126	1.30	32	2046 ±24	2456 ±8	+19	0.1600	0.5	8.240	1.4	0.3734	1.4	0.95
54.1	0.01	107	137	1.32	52	2867 ±27	2736 ±6	-6	0.1893	0.3	14.618	1.2	0.5600	1.2	0.96
56.1	0.05	27	1	0.05	8	1836 ±24	1835 ±21	-0	0.1122	1.1	5.098	1.9	0.3296	1.5	0.80
57.1	0.03	59	114	1.97	21	2219 ±24	2376 ±18	+8	0.1527	1.1	8.650	1.7	0.4109	1.3	0.76

(continued)

Table 1: Summary of SHRIMP U–Pb zircon data for metasediment sample 01-134 (LAB2) *(Continued)*

Grain Spot	% $^{206}Pb_c$	ppm U	ppm Th	$^{232}Th/^{238}U$	(1) ppm $^{206}Pb^*$	(1) $^{206}Pb/^{238}U$ Age (Ma)		(1) $^{207}Pb/^{206}Pb$ Age (Ma)		% Dis-cor-dant	(1) $^{207}Pb^*/^{206}Pb^*$	±%	(1) $^{207}Pb^*/^{235}U$	±%	(1) $^{206}Pb^*/^{238}U$	±%	err corr
60.1	--	61	116	1.96	27	2716	±28	2695	±17	-1	0.1847	1.0	13.341	1.6	0.5239	1.3	0.78
61.1	0.08	19	10	0.52	5	1695	±25	1716	±28	+1	0.1051	1.5	4.357	2.3	0.3007	1.7	0.74
62.1	0.10	11	10	0.89	3	1963	±34	1974	±31	+1	0.1212	1.8	5.948	2.7	0.3559	2.0	0.75
63.1	0.09	74	49	0.69	23	1979	±32	2205	±25	+12	0.1382	1.5	6.847	2.4	0.3593	1.9	0.79
64.1	0.07	55	108	2.04	22	2492	±31	2633	±13	+6	0.1779	0.8	11.572	1.7	0.4719	1.5	0.89
65.1	0.00	189	67	0.37	89	2817	±38	2822	±17	+0	0.1995	1.0	15.075	2.0	0.5480	1.7	0.85
66.1	--	28	46	1.71	6	1337	±18	1742	±27	+26	0.1066	1.5	3.388	2.1	0.2305	1.5	0.72
67.1	0.11	18	1	0.05	4	1465	±23	1566	±37	+7	0.0970	1.9	3.410	2.6	0.2551	1.7	0.67
68.1	--	55	8	0.15	25	2743	±39	2715	±14	-1	0.1869	0.8	13.667	1.9	0.5304	1.7	0.90
69.1	0.16	53	96	1.87	14	1725	±20	2165	±23	+23	0.1351	1.3	5.712	1.8	0.3067	1.3	0.70
70.1	--	78	193	2.55	26	2103	±34	2469	±40	+17	0.1613	2.4	8.575	3.0	0.3857	1.9	0.62
71.1	0.02	65	113	1.81	28	2661	±27	2667	±23	+0	0.1816	1.4	12.791	1.9	0.5109	1.2	0.67
72.1	0.42	19	2	0.10	5	1886	±28	1837	±34	-3	0.1123	1.9	5.262	2.5	0.3398	1.7	0.67
73.1	--	25	32	1.31	7	1758	±24	1775	±23	+1	0.1085	1.3	4.691	2.0	0.3134	1.5	0.77
74.1	--	143	87	0.63	62	2647	±24	2667	±10	+1	0.1815	0.6	12.706	1.3	0.5077	1.1	0.88
75.1	--	65	104	1.67	28	2624	±57	2583	±45	-2	0.1726	2.7	11.954	3.8	0.5023	2.7	0.70
76.1	--	21	33	1.64	6	1813	±26	1773	±30	-3	0.1084	1.6	4.853	2.3	0.3247	1.6	0.70
77.1	0.06	61	109	1.84	27	2683	±28	2687	±15	+0	0.1837	0.9	13.072	1.6	0.5161	1.3	0.83

Errors are 1-sigma; Pb_c and Pb* indicate the common and radiogenic portions, respectively.
Error in Standard calibration was 0.27% (not included in above errors but required when comparing data from different mounts).
(1) Common Pb corrected using measured ^{204}Pb.

lian basement granite bodies of the Long Range, and in particular the Western Brook Pond charnockite (ca. 1466–1499 Ma). These record a pre- to early-Pinwarian age that is best defined in Labrador (1510–1450 Ma). As in the Pinwarian-age rocks of the Long Range, the FdL granite also displays a Grenvillian metamorphic overprint at about 1.0 Ga.

The 966 Ma ages from the granite sample is also consistent with the U–Pb zircon dates from an extensive late Grenvillian granitic suite (Group II) recorded on the Long Range by the Horse Chops and Cloud River granite (999–964 Ma), and the Potato Hill charnockite that were also affected by a younger Grenville metamorphic event at about 970 Ma (Heaman et al. 2002).

The older Mesoproterozoic detrital zircon grains with concordant dates (1750 and 1850 Ma; and up to 2000 Ma) in sample 133 from locality B, are similar to those from pre-Pinwarian crust in the Long Range Inlier basement rocks, (e.g. Labradorian ca. 1650 Ma) and older. Heaman et al. (2002) also identified possible 2140 and 1840 Ma upper intercept ages in relatively old crust in the central portion of the Long Range inlier. Again similar dates are also recorded in the detrital zircon population of the FdL basement gneiss units.

The Neoarchean detrital zircon ages preserved in the FdL paragneiss are more likely to represent a source in the Superior Province, where there is a dominance of Neoarchean rocks preserved directly adjacent to the Grenville province in (e.g. Ludden and Hynes 2000).

In summary, the pre-FdLS basement paragneiss units are consistent with being derived from Precambrian rocks of the Laurentian Shield, while the granite was apparently intruded into pre-Pinwarian sedimentary rocks.

Early Rift Correlative Units

Correlative units of the Lower Fleur de Lys rocks (early rift sequences) in southwestern Newfoundland have yielded a remarkably similar U–Pb age distribution of detrital zircon, also analysed by SHRIMP (Cawood and Nemchin 2001). These can be subdivided into: (1) Archean grains dated between 2890 and 2600 Ma, with a maximum frequency at 2780–2660 Ma, (2) Paleoproterozoic grains ranging from 1950 to 1750 with a pronounced peak at ca. 1850 Ma (with characteristic rounded grains, as also observed in paragneiss of the FdL), (3) Mesoproterozoic to early Neoproterozoic grains dated between 1450 and 950 Ma, and (4) Neoproterozoic igneous zircon ranging between 760 and 570 Ma. Since the youngest grains yield dates of 580–570 Ma this provides a maximum age for accumulation of the FdL sequence here.

Siliciclastic sedimentary rocks of the Bradore Formation, which is deposited directly on the Laurentian basement of the Long Range (including on Belle Isle; Williams and Stevens 1969), have yielded Grenville-age zircon (1223– 930 Ma) but these are also without Late Neoproterozoic zircon (Cawood and Nemchin 2001).

TECTONIC IMPLICATION FOR THE FLEUR DE LYS SUPERGROUP

The FdLS was ductilely deformed during the Taconic orogenic cycle, which

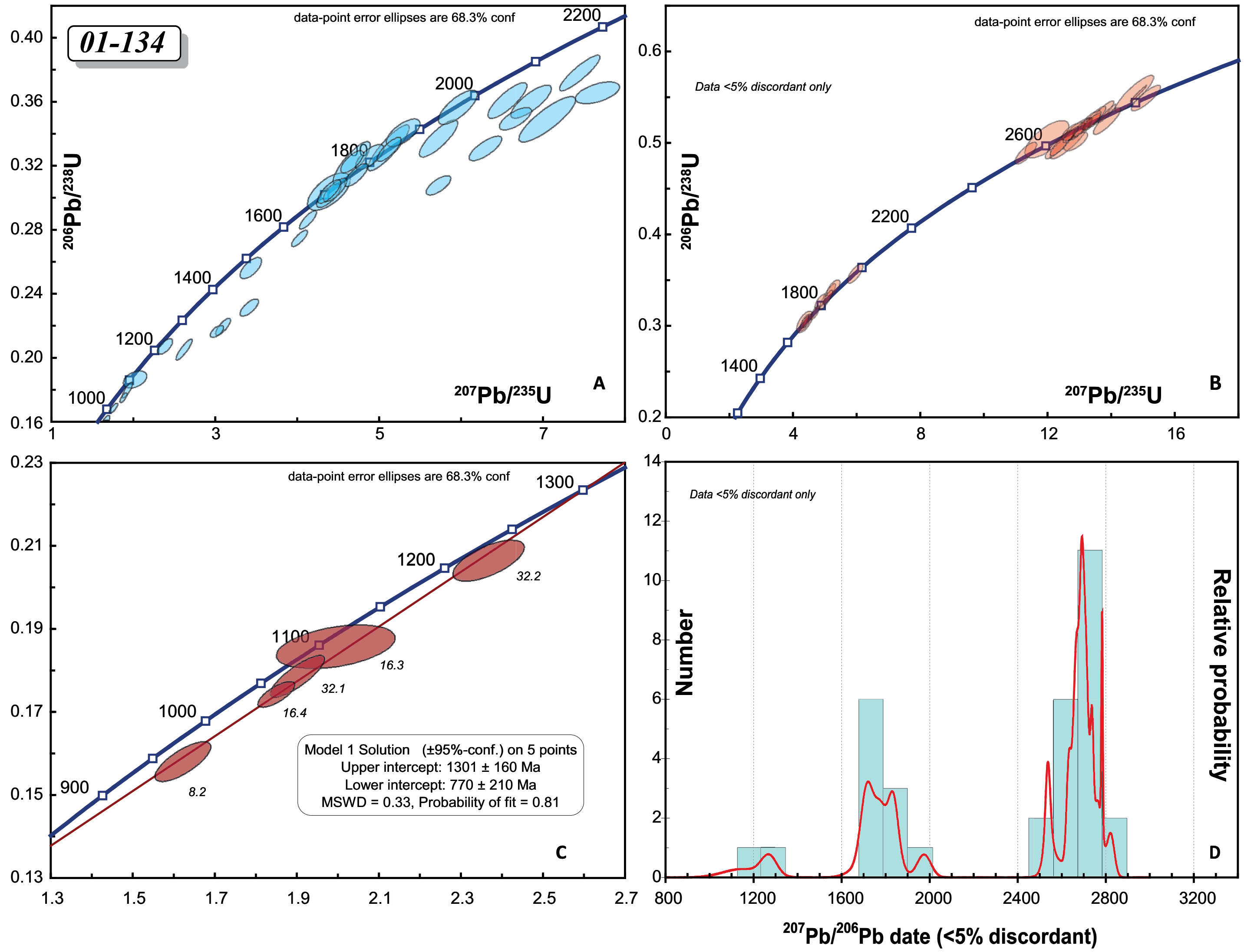

Figure 7. A-C: Concordia plots of zircon from sample 01-134. **D**: Probability density curve showing the major age peaks of identified zircon grains between 1000 and 2900 Ma. See text for further explanations and Table 1 for the data.

comprised three distinct tectonic events between the Late Cambrian and Late Ordovician (500–450 Ma; van Staal et al., 2007). The best preserved event during this period was in the Early to Middle Ordovician (475–459 Ma), related to collision between the Humber margin of Laurentia and the Dashwoods continental fragment with the Notre Dame arc built upon it. This collision represents the first phase towards shaping the present edge of composite Laurentia.

Geologic arguments suggest that the Birchy Complex and correlative sequences represent zones of tectonic mélange that accommodated these initial stages of convergence during the Early to Middle Ordovician by obduction of the Baie Verte Oceanic Track (BVOT) westward across the Humber margin (Bursnall 1975; Hibbard 1983; van Staal et al. 2013). Although such a kinematic model is consistent with Middle Ordovician (ca. 465 Ma) $^{40}Ar/^{39}Ar$ ages of muscovite and hornblende in mafic schist of the Birchy Complex (van Staal et al. 2009a, 2013; Skulski et al. 2010), metamorphic studies suggest that the Birchy Complex and the Rattling Brook allochthon were buried at ca. 479 Ma to significantly greater depths (ca. 33 km) than parts of the structurally overlying BVOT (ca. 20 km). Their present tectonic juxtaposition therefore took place subsequent to peak-Taconic burial metamorphism, and our work here reveals this early tectonism in the Humber Zone, west of the BVL to be Tremadocian in age (ca. 483 Ma), closer to the original date suggested by Church (1969). Taking into consideration the error of this age, the early tectonism might also be Floian in age (the Tremadocian-Floian boundary lies at 478 Ma) in which case the early tectonism would correspond with the estimate of age of the base of the Snooks Arm/Flat Water Pond groups that record emergence of the BVOT at ca. 479 Ma (e.g. Skulski et al. 2010; van Staal et al. 2013), closer to a hornblende date of 477 Ma from a gabbroic amphibolite that intrudes the Birchy Complex of the BVOT (van Staal et al. 2013; C. van Staal pers.

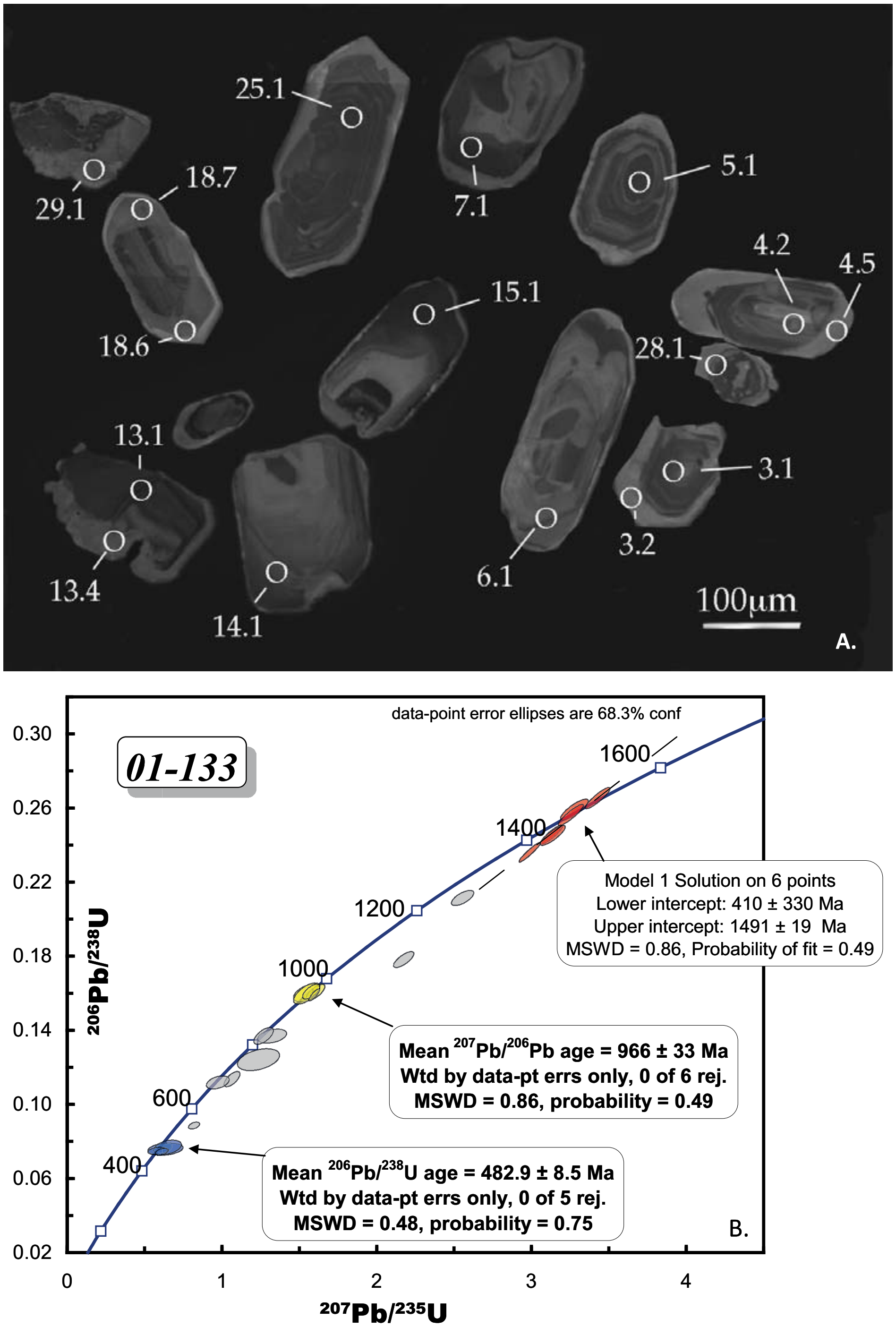

Figure 8. A: Cathodoluminecsence image (CL) of zircon from sample 01-133. Note the narrow outer rims of relatively light-CL and without internal structure, that represent the Ordovician metamorphic overgrowth on the older ca. 1.0-1.5 Ga cores. **B**: Total concordia plot of zircon grains from 01-133. See text for further explanation, and Table 2 for the data.

comm. 2014). Clearly, there is a need for much greater precision in the age determinations of all these rock sequences and their metamorphic minerals.

The closest equivalent Early Ordovician dates in the BVOT relate to the formation of mafic crust for example, the Stog'er Tight gabbro (U–Pb zircon igneous age of 483+3/-2 Ma; Ramezani et al. 2000) and a U–Pb zircon age of ca. 482 Ma on gabbroic amphibolite from the Advocate rocks of the Birchy Complex (with similar geochemistry as the Stog'er Tight gabbro; van Staal et al. 2013). This is consistent with stratigraphic correlation of the BVOT with other Ordovician ophiolitic and volcanic arc/back-arc assemblages of the Dunnage Zone that formed between ca. 489 and 487 Ma (e.g. Skulski et al. 2010). If equivalent ophiolite bodies were emplaced across the FdL terrain at ca. 483 Ma, then the oceanic crust must have been young and relatively warm, consistent with an inverted geotherm from the FdLS into the FdL basement (EPMS) as determined from the metamorphic mineralogy (Jamieson 1990).

It has been suggested that the early high-pressure metamorphism of the FdL formed during A–subduction of parts of the leading edge of the Humber extended margin to eclogite facies depths. This seems a parsimonious model for the present, except there are at least two conundrums. First, the basement rocks were buried deeper during the first metamorphism (Taconic Phase I) by at least 10 km from their overlying FdLS cover. However by the time of the subsequent tectono-metamorphic phase (Taconic II?), their burial depths were similar at ca. 20 km, consistent with their present close proximity and tectono-stratigraphic setting. In turn, this would imply that the basement rocks underthrust the overlying FdL cover between Phase I and Phase II. However, despite the recognition of thrusting high in the FdLS sequences, just to the west of the Baie Verte Line (Kennedy 1971, 1975; Hibbard 1983; van Staal et al. 2013, and references therein), there is no convincing field observation of thrusting or tectonic escape between the FdL conglomerate units and the unconformably underlying basement. This needs renewed field mapping to test.

Second, dating has revealed surprisingly little evidence for early Taconic metamorphism in the FdLS west of the Baie Verte Line (van Staal et al. 2009a). Our work here reveals this early tectonism in the Humber Zone, west of the BVL to be Tremadocian-Floian in age. But, in the present tectonic scheme of things, east-directed subduction began in the Taconic seaway at ca. 490 Ma, forming

Table 2: Summary of SHRIMP U–Pb data for sample 01-133 (LBCR8A)

Grain Spot	% $^{206}Pb_c$	ppm U	ppm Th	$^{232}Th/^{238}U$	(1) ppm $^{206}Pb^*$	(1) $^{206}Pb/^{238}U$ Age (Ma)	(1) $^{207}Pb/^{206}Pb$ Age (Ma)	% Dis-cor-dant	(1) $^{207}Pb^*/^{206}Pb^*$	±%	(1) $^{207}Pb^*/^{235}U$	±%	(1) $^{206}Pb^*/^{238}U$	±%	err corr
1.5	2.04	26	2.1	0.081	2.83	765 ±21	976 ±138	+23	0.0717	6.8	1.244	7.4	0.1260	3.0	0.40
3.1	0.07	237	154	0.67	50.1	1417 ±18	1477 ± 13	+4	0.0925	0.68	3.135	1.5	0.2459	1.4	0.90
3.2	0.08	81	20	0.26	12.4	1063 ±15	1387 ± 26	+23	0.0882	1.4	2.181	2.1	0.1793	1.5	0.75
4.2	0.11	104	77	0.76	23.3	1486 ±19	1466 ± 19	-1	0.0919	10	3.286	1.8	0.2592	1.5	0.83
4.5	0.41	31	2.3	0.077	2.10	489 ±11	516 ±127	+6	0.0576	5.8	0.626	6.2	0.0788	2.4	0.38
5.1	0.00	457	208	0.47	92.9	1368 ±17	1465 ± 9	+7	0.0919	0.47	2.996	1.5	0.2365	1.4	0.95
6.1	0.19	153	86	0.58	21.3	968 ±15	993 ± 34	+2	0.0722	1.7	1.615	2.3	0.1621	1.6	0.69
7.1	0.00	272	83	0.32	37.5	960 ±12	938 ± 17	-2	0.0703	0.84	1.557	1.6	0.1605	1.4	0.85
8.1	0.04	326	374	1.18	71.9	1472 ±19	1475 ± 13	0	0.0924	0.69	3.266	1.6	0.2565	1.5	0.90
9.1	0.41	36	26	0.73	5.05	966 ±16	919 ± 62	-5	0.0697	3.0	1.553	3.5	0.1616	1.8	0.51
9.2	0.53	81	25	0.32	9.75	836 ±13	924 ± 86	+9	0.0699	4.2	1.334	4.5	0.1385	1.7	0.37
10.1	0.05	148	47	0.33	31.2	1418 ±18	1484 ± 17	+4	0.0928	0.88	3.150	1.7	0.2461	1.4	0.85
11.1	0.26	65	53	0.84	9.05	963 ±20	930 ± 54	-4	0.0701	2.7	1.557	3.5	0.1612	2.3	0.65
12.1	0.00	42	13	0.31	4.18	706 ±15	845 ± 53	+17	0.0672	2.5	1.073	3.4	0.1157	2.3	0.67
13.1	-0.06	267	103	0.40	60.7	1517 ±18	1508 ± 12	-1	0.0940	0.65	3.439	1.5	0.2653	1.4	0.90
13.4	2.38	33	1.0	0.031	2.23	486 ±13	578 ±252	+16	0.0593	11.6	0.641	11.9	0.0784	2.8	0.24
14.1	0.00	95	48	0.52	13.3	969 ±14	951 ± 35	-2	0.0708	1.7	1.583	2.3	0.1623	1.6	0.67
15.1	0.04	307	107	0.36	42.4	961 ±12	996 ± 17	+4	0.0724	0.83	1.604	1.6	0.1607	1.4	0.86
18.6	1.22	38	1.9	0.052	2.65	499 ±15	543 ±181	+8	0.0583	8.3	0.647	8.9	0.0804	3.2	0.36
18.7	0.34	41	0.1	0.002	2.76	480 ±7	426 ±138	-13	0.0554	6.2	0.590	6.4	0.0773	1.6	0.25
21.3	0.19	32	0.1	0.003	2.12	478 ±8	535 ± 81	+11	0.0581	3.7	0.617	4.1	0.0770	1.7	0.41
22.4	1.08	100	36.2	0.37	18.23	1239 ±14	1374 ± 27	+11	0.0876	1.4	2.560	1.8	0.2120	1.2	0.67
28.1	1.00	58	2.9	0.052	4.54	561 ± 7	796 ± 59	+31	0.0657	2.8	0.823	3.1	0.0909	1.4	0.44
29.1	0.44	108	1.3	0.013	12.85	837 ±16	833 ± 51	0	0.0669	2.4	1.278	3.2	0.1387	2.0	0.63
30.1	0.68	50	15.6	0.33	4.85	695 ±14	697 ± 97	0	0.0627	4.6	0.984	5.0	0.1138	2.1	0.42

Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.
Error in Standard calibration was 0.34% (not included in above errors but required when comparing data from different mounts).
(1) Common Pb corrected using measured ^{204}Pb.

the 490–483 Ma supra-subduction zone BVOT (van Staal et al. 2013, and references therein). van Staal et al. (2013) hinted also that earlier west-directed subduction within the Taconic seaway may have started during the Middle Cambrian, and the metamorphic ages preserved in rims on the old zircon grains in the FdL basement likely represent this early event. The reason for this Late Cambrian (495–490 Ma) subduction in the Taconic Seaway (as evidently represented by the high-pressure metamorphism in the FdL), rather than in Iapetus, remains elusive (van Staal et al. 2013).

CONCLUSION

A FdLS cover of Late Neoproterozoic coarse siliciclastic rocks, with detrital zircon ranging between ca. 2900 and 570 (possibly 550) Ma — with its lower sequences intruded by basalt (ca. 615 Ma), and its upper sequences overlain by Lower Cambrian fine siliciclastic rocks and marble — unconformably overlies a complex Laurentian basement. The visionary correlation of similar sequences on Belle Isle, across to the Burlington Peninsula by Williams and Stevens some 45 years ago, appears vindicated.

Both FdL cover and basement on the Burlington Peninsula were intensely deformed and metamorphosed during the formation of the early Paleozoic Appalachian orogen. Early high pressure – low temperature metamorphic conditions of the basement peaked around 483 Ma, at a depth more than 10 km in excess of its overlying FdLS cover, but at the time of the breakdown of eclogite (ca. 465 Ma), differential pressure had almost dissipated as lower pressure – higher temperature conditions of basement and cover converged at mid-crustal depths of about 20 km. Thereafter the successions were tectonically exhumed in unison. Unfortunately variable reported Ar/Ar dates on similar metamorphic mineral assemblages can be used to support exhumation rates during either extraordinary rapid cooling or very slow cooling by 465 Ma or 420 Ma, respectively; neither seem solid on geological grounds.

Thus, the internal zone of the Humber margin of the Appalachians in Newfoundland is in dire need of a well-constructed program of remapping with more precise geochronometry. It will be interesting to verify the age of deformation of the FdL and basement on Belle Isle, and it would be instructive to carry out a geo- and thermo-chronological study on the FdLS conglomerate clasts, best exposed along Middle Arm, to compare and contrast with those in the FdL on Belle Isle; and to construct a more precise tectonothermal history from rifting to tectonic burial and subsequent exhumation across the Humber Margin. Using the relatively undeformed Fleur de Lys and its basement on Belle Isle as a new base may considerably strengthen the geo-cocktails first served by Williams and Stevens (1969). Clearly what Williams started in the 1960s has not yet been completed, and he would be the first to endorse further work with prospects of an improved well-aged vintage.

ACKNOWLEDGEMENTS

This contribution was spurred on by Brendan Murphy, who never stopped with subtle enquiries about progress, and without which it would never have seen the light of day. We thank Cees van Staal and Tom Skulski for updating us on recent work in NE Newfoundland, and for their permission to modify Figure 1 from their maps; and to Cees for his generosity in sharing unpublished eclogite data and ideas.

The advances in dating techniques over the last 25 years or so surely have improved field interpretations, and we would like to acknowledge the role of many dedicated geochronologists for collaborative fun; we are sure that Hank Williams and John Dewey would endorse this acknowledgement. Our hope now is that new integrated projects on the Burlington Peninsula will continue to modify the present models of this unique Appalachian Laboratory. There are no doubt surprises in store.

MdW would like to express his thanks to John Dewey for always urging that field mapping must surely form the very foundations of our knowledge of how the earth works and especially to test how it worked over deep time; to Hank Williams who shared this wisdom over nearly half a century; and to Cees van Staal who now truly has taken over their batons. He would also like to acknowledge the Newfoundland generosity and friendship at the Pinkson family home, his field abode in Seal Cove; Blanche and Frank have both passed away, but correspondence with their (grand)children is still alive and well after nearly 45 years. Living in Bear Cove with fisherman and brewer Arthur Gavin (drowned in 1976) enriched his FdL memories forever.

We are grateful for critical comments by Cees van Staal and Paul Hoffman that helped improve this paper; and to Cindy Murphy for generous help and patience with our disruptive submission process. This is AEON contribution number 122.

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Received October 2013
Accepted as revised March 2014