Stratigraphic evolution of the Neoproterozoic Callison Lake Formation: Linking the break-up of Rodinia to the Islay carbon isotope excursion

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1 **STRATIGRAPHIC EVOLUTION OF THE NEOPROTEROZOIC CALLISON**
2 **LAKE FORMATION: LINKING THE BREAK-UP OF RODINIA TO THE ISLAY**
3 **CARBON ISOTOPE EXCURSION**

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19 **ABSTRACT. The ~780-540 Ma Windermere Supergroup of western North America**
20 **records the protracted development of the western Laurentian passive margin and**
21 **provides insights into the nature, timing, and kinematics of Rodinia's fragmentation.**
22 **Here we present a refined tectono- and chemo-stratigraphic model for circa 780–720**
23 **Ma sedimentation in NW Canada through a study of the Callison Lake Formation**

(formalized herein) of the Mount Harper Group, spectacularly exposed in the Coal Creek and Hart River inliers of the Ogilvie Mountains of Yukon, Canada. Twenty-one stratigraphic sections are integrated with geological mapping, facies analysis, carbon and oxygen isotope chemostratigraphy, and Re-Os geochronology to provide a depositional reconstruction for the Callison Lake Formation. Mixed siliciclastic, carbonate, and evaporite sediments accumulated in marginal marine embayments formed in discrete hangingwall depocenters of a prominent Windermere extensional fault zone. Deposition of the Windermere Supergroup in NW Canada post dates the eruption of the circa 780 Ma Gunbarrel Large Igneous Province by ~30 million years, is locally associated with compressional or transpressional tectonism, and predates the successful rift-drift transition by ~200 million years. In order to accommodate evidence for coeval extensional and compressional tectonism, abrupt facies change, and Neoproterozoic fault geometries, we propose that NW Laurentia experienced strike-slip deformation during the ~740–660 Ma early fragmentation of the supercontinent Rodinia. Sequence stratigraphic data from the Callison Lake Formation and other basal Windermere successions in the northern Canadian Cordillera delineate three distinct depositional sequences, or transgressive-regressive (T-R) cycles, that are coeval with similar stratigraphic packages in the ~780–720 Ma Chuar-Uinta Mountain-Pahrump basins of the western United States. The global circa 735 Ma Islay carbon isotope excursion is consistently present in carbonate strata of the third T-R cycle and is interpreted to represent a primary perturbation to the global carbon cycle, possibly driven by the uplift and weathering of extensive shallow epicontinental seaways and evaporite basins.

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48 Key words: Windermere Supergroup, Neoproterozoic chemostratigraphy, Islay carbon
49 isotope excursion, Mount Harper Group, Callison Lake Formation

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INTRODUCTION

52

53 Neoproterozoic sedimentary deposits of western North America provide a critical
54 record of the protracted breakup of the supercontinent Rodinia (Stewart, 1972; Young
55 and others, 1979; Ross, 1991; Yonkee and others, 2014) and form the backbone to many
56 global geochemical, paleontological, and geochronological compilations (Narbonne and
57 others, 1994; Halverson and others, 2005; Rooney and others, 2015, Cohen and
58 Macdonald, 2015). The tectonic setting and evolution of these sedimentary successions
59 also provide essential geological context for Neoproterozoic climate change. For instance,
60 Rodinia's fragmentation at low latitude may have set the stage for runaway global
61 cooling that resulted in global Cryogenian glaciation (Kirshvink, 1992; Hoffman and
62 others, 1998; Hoffman and Schrag, 2002; Schrag and others, 2002). Recent
63 interpretations for the initiation of these glaciations point to the consumption of CO₂ and
64 O₂ via uplift and weathering of extensive continental flood basalts, many of which were
65 centered along the Neoproterozoic margins of Laurentia (Goddris and others, 2003;
66 Donnadieu and others, 2004; Macdonald and others, 2010; Halverson and others, 2014;
67 Rooney and others, 2014). Furthermore, high-precision U-Pb chemical abrasion-isotope
68 dilution-thermal ionization mass spectrometry (CA-ID-TIMS) zircon ages on volcanic
69 tuffs and Re-Os organic-rich rock (ORR) ages on black shale and carbonate interbedded
 with Neoproterozoic marine strata from NW Canada have recently established firm

temporal constraints on critical events in Neoproterozoic Earth history, including the onset of the Sturtian glaciation and its relationship to the global Islay carbon isotope excursion (ICIE; also called the “Islay anomaly”) (Macdonald and others, 2010; Rooney and others, 2014; Strauss and others, 2014a). Yet, many aspects of the early tectonic and environmental evolution of the western Laurentian margin remain controversial, such as the temporal and spatial record of extension, the exact timing of rift-drift transition, and the paleogeographic arrangement of circum-Laurentian continents in Rodinia.

Building on the stratigraphic compilations of Gabrielse (1967) and Crittenden and others (1971), Stewart (1972) was the first Cordilleran geologist to highlight the “Windermere Group” and equivalent rocks as recording the rift phase of the western Laurentian passive margin. The <6 km thick Windermere Supergroup encompasses ca. 780–540 Ma rocks of the Canadian Cordillera (Sequence C of Young and others, 1979), but stratigraphic equivalents of these deposits span the length of North America from Mexico to Alaska (Wheeler and McFeely, 1991). The concept of Neoproterozoic extensional opening of the paleo-Pacific Ocean persisted for over a decade until papers using backstripping techniques to generate subsidence curves for the early Paleozoic passive margin indicated that the rift-drift transition occurred at ~540 Ma, ~200 million years after the major expression of Windermere Supergroup rift-related sedimentation and volcanism at ~780–680 Ma (Armin and Mayer, 1983; Bond and others, 1983; Bond and Kominiz, 1984; Devlin and Bond, 1988; Levy and Christie-Blick, 1989). Therefore, most recent reconstructions (for example, Colpron and others, 2002) indicate a protracted two-stage rift history for the western margin of Laurentia (ca. 780–700 and ca. 575–540

Ma), with localized pulses of extension persisting into the early Paleozoic (for example, Goodfellow and others, 1995; Cecile and others, 1997; Pyle and Barnes, 2003).

Mid-Neoproterozoic (Tonian–Cryogenian) Laurentian basins provide a particularly important record of the nature, timing, and driving mechanisms for Rodinia’s initial fragmentation and inform the development of a viable kinematic model for Cordillera-wide extension. The evolution of these geographically isolated sedimentary deposits has been attributed to subsidence associated with mantle plume activity and emplacement of the ca. 780 Ma Gunbarrel Large Igneous Province (LIP) (Li and others, 1999; 2008; Harlan and others, 2003; Macdonald and others, 2012; Sandeman and others, 2014; Yonkee and others, 2014). In the western United States, these strata include the Chuar Group (Gp) of Arizona, the Pahrump Gp of California, and the Uinta Mountains Gp and Big Cottonwood Formation (Fm) of Utah, which are collectively referred to as the ChUMP basins (Dehler and others, 2010 and references therein). Equivalent strata in NW Canada, which are the focus of this study, include the Coates Lake Gp of the Mackenzie Mountains, Northwest Territories (NWT), and the Mount Harper Gp of the Ogilvie Mountains, Yukon (for example, Jefferson and Parrish, 1989; Mustard and Roots, 1997; Macdonald and others, 2012). In this contribution, we refine the record of early Windermere Supergroup sedimentation in the Ogilvie Mountains of Yukon, Canada, through an integrated study of the Callison Lake Fm (previously called the Callison Lake dolostone; table 1) of the Mount Harper Gp. We present new geological mapping combined with detailed measured stratigraphic sections, sequence stratigraphy, and carbon and oxygen isotope chemostratigraphy to develop a depositional model for the

Callison Lake Fm, provide an updated interpretation for mid-Neoproterozoic extension in NW Canada, and assess the origin of the ICIE.

GEOLOGICAL BACKGROUND AND PREVIOUS WORK

Proterozoic sedimentary strata in northwest Canada are discontinuously exposed throughout the Cordilleran fold and thrust belt and in erosional inliers through Phanerozoic strata (fig. 1A). Although crystalline basement is not exposed in Yukon, Proterozoic sedimentary successions further to the east in the NWT rest unconformably on the ca. 1880–1840 Ma western Bear Province of the Wopmay Orogen (Hoffman, 1989; Villeneuve and others, 1991; Bowring and Grotzinger, 1992). Young and coworkers (1979) divided the Proterozoic sedimentary successions of NW Canada into three discrete unconformity bounded “sequences” (fig. 1): Sequence A (~1.7–1.2 Ga Wernecke Supergroup and equivalents), Sequence B (~1.2–0.78 Ga Mackenzie Mountains Supergroup and equivalents), and Sequence C (~0.78–0.54 Ga Windermere Supergroup and equivalents). Over the past three decades, this classification has been refined and subdivided following the recognition of regional unconformities, unique basin forming events, and new radiometric age constraints (Cook and Maclean, 1995; Thorkelson and others, 2005; Turner and others, 2011; Macdonald and others, 2012; Furlanetto and others, 2013; Medig and others, 2014; Thomson and others, 2014; 2015a).

The Callison Lake Fm of the Mount Harper Gp (Windermere Supergroup) is part of the Mackenzie-Ogilvie Platform, a paleogeographic feature of the Foreland Belt of the Canadian Cordillera (Gordey and Anderson, 1993; Cecile and others, 1997; Norris, 1997). More specifically, these strata are exposed in the Coal Creek and Hart River inliers of the

Yukon Block (see Yukon Stable Block of Jeletsky, 1962), an independent lithospheric block that underlies most of north-central Yukon from the Alaskan border to the Richardson Fault Array near the Yukon-NWT border (fig. 1; Jeletsky, 1962; Lenz, 1972; Roots and Thompson, 1992; Abbott, 1997; Cecile and others, 1997; Morrow, 1999; Thorkelson and others, 2005). The Dawson Fault (also known as the Dawson Thrust) and Richardson Fault Array bound the Yukon Block to the south and east, respectively (fig. 1); both of these long-lived structural features delineate a platform-to-basin transition – the Richardson Trough to the east and the Selwyn Basin to the south (for example, Gordey and Anderson, 1993; Abbott, 1997; Cecile and others, 1997). The northern and western boundaries of the Yukon Block are not as clearly defined.

Proterozoic strata of the Yukon Block are superbly exposed in the Tatonduk, Coal Creek, Hart River, and Wernecke inliers of Yukon, all of which share a similar, yet distinct, depositional history. Despite synsedimentary tectonism and associated complex facies change (for example, Aitken, 1981; 1991; Eisbacher, 1981; 1985; Jefferson, ms, 1983; Jefferson and Parrish, 1989; Abbott, 1997; Thorkelson and others, 2005; Turner and Long, 2008; Turner and others, 2011; Macdonald and others, 2012), recent work has begun to shed light on interregional correlations among the Proterozoic inliers of Yukon and strengthen connections to the well documented successions of the Mackenzie Mountains and Amundsen Basin (fig. 2; Rainbird and others, 1996; Abbott, 1997; Thorkelson and others, 2005; Macdonald and Roots, 2010; Macdonald and others 2010; 2011; 2012; Medig and others, 2010; 2014; Turner, 2011; Turner and others, 2011; Halverson and others, 2012; Cox and others, 2013; Furlanetto and others, 2013; van Acken and others, 2013; Kunzmann and others, 2014; Thomson and others, 2014; 2015a).

For the purposes of this contribution, we will not discuss the geology of the Tatonduk inlier any further because basal Windermere Supergroup strata are not well developed and exposed in this region (Macdonald and others, 2011). Importantly, there is a clear difference in equivalent Proterozoic strata across the Richardson Fault Array, particularly the Snake River and Knorr faults mapped by Aitken and Cook (1974), Eisbacher (1977; 1981), and Norris (1982). Stratigraphic correlation of Yukon Block inliers to the adjacent Mackenzie Mountains requires significant facies change and some component(s) of tectonic rotation and strike-slip displacement along the ancestral Richardson Fault Array (figs. 1, 2; Wheeler, 1954; Gabrielse, 1967; Aitken and Cook, 1974; Eisbacher, 1977; 1978; 1981; Bell, 1982; Jefferson, ms, 1983; Jefferson and Parrish, 1989; Aitken and McMechan, 1992; Park and others, 1992; Abbott, 1997; Thorkelson, 2000; Thorkelson and others, 2005).

Hart River Inlier

The Hart River inlier is located in the southeastern Ogilvie Mountains (figs. 1, 2, 3) and contains Proterozoic strata of the Wernecke, Pinguicula, Fifteenmile, Rapitan, and Mount Harper groups. Green's (1972) regional mapping of the Dawson (116B and C), Larsen Creek (116A), and Nash Creek (106D) map areas established the regional stratigraphic and structural framework for the Ogilvie Mountains, upon which Abbott (1993; 1997) added detailed mapping of the Mark Creek area (fig. 3). The oldest strata exposed in the Hart River inlier comprise mixed siliciclastic and carbonate strata of the Quartet and Gillespie Lake groups (Wernecke Supergroup), which were most likely deposited in a Mesoproterozoic intracratonic basin or along a poorly understood passive

continental margin (Green, 1972; Abbott, 1993; 1997; Delaney and others, 1981; Thorkelson and others, 2000; 2005; Furlanetto and others, 2013). The Hart River basalts unconformably overlie the Wernecke Supergroup and are assumed to be extrusive equivalents of the circa 1380 Ma Hart River sills (Abbott, 1997). These volcanic rocks are unconformably overlain by ~2.5 km of mixed siliciclastic-carbonate strata of the Pinguicula and Fifteenmile groups, which consist of two separate successions defined by map units PPA–C and PPD1–3 and separated by a prominent angular unconformity (figs. 2, 3; Abbott, 1997). Halverson and others (2012), following Medig and others (2010), correlated the upper Pinguicula Gp of the Hart River inlier (map units PPD1-3) with the informal Gibben and Chandindu formations of the lower Fifteenmile Gp of the Coal Creek inlier and the Hematite Creek Gp of the Wernecke inlier (figs. 2, 3).

In the Hart River inlier, the Callison Lake Fm rests with an angular unconformity on Pinguicula, Wernecke, and lower Fifteenmile Gp strata, suggestive of significant tectonic uplift and erosion prior to deposition (fig. 3; Abbott, 1997; Macdonald and Roots, 2010; Macdonald and others, 2010; Halverson and others, 2012). Abbott (1993; 1997) was the first to recognize the Callison Lake Fm and described it as a well-bedded, light gray weathering dolomite unit characterized by stromatolites, pisolites, intraformational breccias, “cryptalgal” laminations, and abundant chert lenses. No stratigraphic sections of the Callison Lake Fm were measured, and Abbott (1997) tentatively correlated these strata with the upper Fifteenmile Gp (map unit PF1) of the Coal Creek inlier and the Little Dal Gp of the Mackenzie Mountains. Based on similarities in lithology and stratigraphic position, Macdonald and Roots (2010) subsequently correlated the Callison Lake Fm with the upper Fifteenmile Gp of the Coal Creek inlier (map units PF2 and PF3).

The Callison Lake Fm is disconformably or erosionally overlain by conglomerate, glacial diamictite, or mixed carbonate-siliciclastic strata of the Mount Harper, Rapitan, Hay Creek, and ‘upper’ groups (figs. 2, 3; Abbott, 1997; Macdonald and Roots, 2010; Macdonald and others, 2010).

Coal Creek Inlier

The Coal Creek inlier is located in the south-central Ogilvie Mountains (figs. 1, 2, 4) and contains Proterozoic strata of the Wernecke, Pinguicula, Fifteenmile, Rapitan, Mount Harper, Hay Creek, and “upper” groups. Following Green’s (1972) establishment of the regional geologic framework for the Dawson (116B and C) map sheet, subsequent detailed mapping and stratigraphic projects by Thompson and Roots (1982), Blaise and Mercier (1984), Thompson and others (1987; 1994), Roots (ms, 1987), Mercier (1989), Mustard (ms, 1990; 1991), and Mustard and Roots (1997) subdivided the Proterozoic stratigraphic succession and established the informal Mount Harper and Fifteenmile groups. The oldest rocks exposed in the Coal Creek inlier comprise mixed siliciclastic and carbonate strata of the Fairchild Lake, Quartet, and Gillespie Lake groups (Wernecke Supergroup) (fig. 4). The overlying Fifteenmile Gp was originally subdivided into lower (PR1–5) and upper (PF1–3) subgroups with five and three map units (Thompson and others, 1994), respectively, that locally unconformably overlie the Pinguicula Gp and Wernecke Supergroup. Macdonald and Roots (2010) and Macdonald and others (2011; 2012) recently revised the Fifteenmile Gp sedimentary succession and discarded many informal map units used by Thompson and others (1994). According to this new scheme, the Fifteenmile Gp can be subdivided into an ~1 km thick “Reefal assemblage” (previous

map units PR1–5 and PF1a) of mixed shale and carbonate strata. Zircon extracted from a quartz-phyric tuff interbedded with shale in the upper portion of the Reefal assemblage constrains its depositional age to 811.51 ± 0.25 Ma (CA-ID-TIMS; Macdonald and others, 2010). Halverson and others (2012) proposed a tripartite subdivision of the Fifteenmile Gp into the Gibben and Chandindu formations and the Reefal assemblage to clarify correlations among Proterozoic inliers, including the relationship between the Fifteenmile and Pinguicula groups. Macdonald and others (2012) and Kunzmann and others (2014) expanded upon this by providing a detailed description of the basin forming mechanism and regional correlations of these strata. Lower Fifteenmile Gp strata are gradationally overlain by the “Craggy dolostone” (informal term; formerly map unit PF1), an ~500 m thick, massive and highly silicified dolostone unit (Thompson and others, 1994; Macdonald and others, 2012). The Craggy dolostone is overlain by the Callison Lake Fm (formerly map units PF2 and PF3; table 1). Although this contact was previously considered gradational (Thompson and others, 1994; Mustard and Roots, 1997), here we document evidence for subaerial exposure, paleokarst development, and angular stratigraphic discordance along this contact (fig. 5A-C). The recognition in both the Coal Creek and Hart River inliers of a significant exposure surface and angular unconformity separating the Callison Lake Fm from the underlying Fifteenmile Gp (Abbott, 1997) suggests that the Callison Lake Fm is more closely related to the overlying Mount Harper Gp, which forms the base of the Windermere Supergroup (Macdonald and Roots, 2010; Macdonald and others, 2011; Strauss and others, 2014a).

STRATIGRAPHIC FRAMEWORK AND FORMALIZATION OF THE MOUNT HARPER GROUP

The Mount Harper Gp of the Coal Creek inlier consists of three separate units (fig. 2): the ~500 m thick mixed siliciclastic-carbonate Callison Lake Fm that is described herein (table 1), an ~1100 m thick rift-related siliciclastic succession called the Seela Pass Fm (previously termed the “lower Mount Harper Group” or “Mount Harper conglomerate” but formalized herein; table 2), and an ~1200 m thick bimodal volcanic suite called the Mount Harper volcanics (previously termed the “Mount Harper Volcanic Complex”) (Strauss and others, 2014b and references therein). The overlying ~300–1400 m thick siliciclastic and carbonate succession, previously referred to as the “upper Mount Harper Group,” has been correlated with distinct portions of the Rapitan, Hay Creek, and “upper” groups of the Mackenzie Mountains and has been abandoned (Macdonald and others, 2011; Strauss and others, 2014b). The Mount Harper Gp stratigraphic succession in the Hart River inlier is not as well characterized (Abbott, 1997); however, our preliminary geological mapping differentiates the Mount Harper Gp from the overlying Rapitan, Hay Creek, and “upper” groups (fig. 4).

The Seela Pass Fm (table 2) is composed of up to ~1100 m of continental strata intimately associated with a syndepositional, north-side-down normal fault (fig. 5E; Thompson and others, 1987; Mustard, ms, 1990; Mustard, 1991; Mustard and Donaldson, 1990; Mustard and Roots, 1997; Strauss and others, 2014b). Mustard and Donaldson (1990) and Mustard (1991) describe the contact of the Seela Pass and Callison Lake formations as a distinct disconformity marked by up to ~100 m of paleorelief and thick paleokarst breccias. A thick tapering wedge of fault-derived breccia (Mustard, 1991),

distinct from the laterally equivalent karst breccia, is preserved along the down-dropped side of the Harper Fault (Mustard, ms, 1990; Mustard, 1991; Mustard and Donaldson, 1990; Mustard and Roots, 1997). This geographically isolated wedge grades both northward and eastward into massive and disorganized beds of coarse-grained boulder to gravel conglomerate interpreted as upper- to mid-fan alluvial fan debris flows (Mustard, 1991). Farther to the northeast, alluvial fan deposits of the Seela Pass Fm grade laterally into conglomerate, sandstone, and mudstone of distal fan delta deposits that yield northward directed paleocurrent directions (Mustard, 1991). This maroon mudstone facies assemblage records a distinct coarsening-upwards package, which Mustard (1991) interpreted as evidence for progradation of Seela Pass alluvial fans into a lacustrine or marginal marine setting. Clastic deposition of the Seela Pass Fm ceased with the onset of Mount Harper volcanism (Mustard, 1991; Mustard and Roots, 1997).

The ancestral Harper Fault dips 50-60° northward and delineates the southernmost exposure of geographically isolated outcrops of the Seela Pass and Callison Lake formations in the Coal Creek inlier (figs. 4, 5E; Mustard and Roots, 1997). Detailed mapping suggests these limited regional exposures are of primary depositional origin, outlining the remnants of an ~10 (N-S) X ~80-100 (E-W) km Proterozoic half graben (Roots, ms, 1987; Mustard, 1991; Mustard and Roots, 1997; Strauss and others, 2014b). The footwall of the Harper Fault is composed of crenulated quartzite and shale of the Quartet Gp and tectonized carbonate of the Gillespie Lake Gp (Wernecke Supergroup) (fig. 4; Thompson and others, 1994; Mustard and Roots, 1997; Strauss and others, 2014b), providing a maximum ~4 km vertical offset assuming the sub-Mount Harper Gp succession in the hanging wall is of similar thickness to the eroded remnants of the

footwall (Mustard, 1991). A minimum offset is ~1100 m, as indicated by the composite measured thickness of the Seela Pass Fm directly north of the Harper Fault (Mustard, 1991). Evidence for maximum progressive downward denudation of the footwall fault scarp is supported by inverted stratigraphy in fault-proximal Seela Pass conglomerates (Mustard, 1991; Macdonald and others, 2011).

The bimodal Mount Harper volcanics are divided into six informal units and two distinct compositional suites (fig. 4; Roots, ms, 1987; Mustard and Roots, 1997; Cox and others, 2013). The basal volcanic suite, members A–C, forms an ~1200 m thick basaltic edifice that conformably overlies Seela Pass strata in the hangingwall of the Harper Fault and unconformably overlies Quartet and Gillespie Lake strata in the footwall, attesting to uplift and erosion of the footwall block prior to volcanism (Mustard and Roots, 1997). These mafic lavas formed in both subaerial and subaqueous settings and most likely initially erupted subaqueously onto damp substrate of the Seela Pass Fm, judging from reworked volcanoclastics, convolute internal stratification, entrained clasts in lowermost lava flows, and loading structures in the uppermost Seela Pass Fm (Roots, ms, 1987; Mustard, 1991; Mustard and Roots, 1997; Cox and others, 2013). The upper suite is characterized by intermediate andesites and rhyolites of members D–F that were dated by U-Pb CA-ID-TIMS on zircon at 717.43 ± 0.14 Ma (Macdonald and others, 2010). The Mount Harper Volcanics are overlain conformably by, and locally interfinger with, conglomerate, sandstone, and glacial diamictite of the informal Eagle Creek formation (Strauss and others, 2014b) of the Rapitan Gp (Macdonald and others, 2010; 2011; Strauss and others, 2014b).

As noted above, Abbott (1993; 1997) was the first to recognize the Callison Lake Fm in the Hart River inlier, but no measured sections were presented in the literature until Macdonald and Roots (2010) and Macdonald and others (2010; 2011) presented preliminary coarse stratigraphic descriptions and carbon isotope chemostratigraphy from one section in the Coal Creek inlier. Subsequently, Tosca and others (2011) described the mineralogy of unusual sedimentary talc deposits from the lower Callison Lake Fm and Strauss and others (2014a) reported 739.9 ± 6.1 Ma vase-shaped microfossil (VSM) assemblages from black shale in upper Callison Lake strata (fig. 2). More recently, Rooney and others (2015) presented another Re-Os ORR age of 752.7 ± 5.5 Ma from the lower Callison Lake Fm (fig. 2). Here, we place all of these previous descriptions and geochronological constraints into a refined stratigraphic context while formalizing the Callison Lake Fm (table 1).

METHODS

Geological mapping was undertaken in the Coal Creek inlier over two- to four-week summer field seasons from 2009–2013 and in the Hart River inlier during the summers of 2009, 2011, and 2012. Twenty-one detailed stratigraphic sections of the Callison Lake Fm were logged by measuring stick, tape measure, and Jacob staff at m- to cm-scale during mapping projects from 2009–2013 (coordinates of logged sections are provided in table DR1 of the AJS supplementary data file¹). Owing to the constraint of available exposure, some of these measured sections are composite (figs. 6, 7). The

¹All GPS coordinates for stratigraphic sections (table DR1) and carbon and oxygen isotope data (table DR2) are presented online as a supplementary electronic data file (<http://earth.geology.yale.edu/~ajs/SupplementaryData/xxx/xxx>).

measured sections are divided into siliciclastic, carbonate, and diagenetic lithofacies based on composition, texture, bedding style, and sedimentary structures (table 3). Paleocurrent analysis was not performed due to the general paucity of reliable current indicators.

We present 1222 new carbonate carbon ($\delta^{13}\text{C}_{\text{carb}}$) and oxygen ($\delta^{18}\text{O}_{\text{carb}}$) isotopic measurements from specified stratigraphic sections (all raw data are presented in table DR2 of the AJS supplementary data file¹). Fist- to golf ball-sized hand samples were collected at 0.5–2 m resolution through measured sections for carbonate carbon and oxygen isotope chemostratigraphy. $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ isotopic results are reported in per mil notation of $^{13}\text{C}/^{12}\text{C}$ and $^{18}\text{O}/^{16}\text{O}$, respectively, relative to the standard VPDB (Vienna-Pee-Dee Belemnite). Carbonate samples were cut perpendicular to bedding and primary lithofacies were carefully microdilled (~2-10 mg of powder) to avoid secondary veins, cements, and siliciclastic components. $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ isotopic data were acquired simultaneously on a VG Optima dual inlet isotope ratio mass spectrometer coupled with a VG Isocarb preparation device (Micromass, Milford, MA) in the Laboratory for Geochemical Oceanography at Harvard University. Approximately 1 mg of sample powder was reacted in a common, purified phosphoric acid (H_3PO_4) bath at 90°C. The evolved CO_2 was collected cryogenically and analyzed using an in-house reference gas. Measured data were calibrated to VPDB using the Cararra marble standard (CM2). Total analytical errors (1σ) are better than $\pm 0.1\text{‰}$ for both $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ based on repeat analysis of standards and samples. Increasing the reaction time to eleven minutes for dolomite samples minimized “memory effects” resulting from the common

acid bath system, with the total memory effect estimated at <0.1‰ based on reproducibility of standards run directly after samples.

SEDIMENTOLOGY AND STRATIGRAPHY OF THE CALLISON LAKE FORMATION

The Callison Lake Fm is divided here into four informal members (Heterolithic, Talc, Ramp, and Transitional) that can be correlated between the Coal Creek and Hart River inliers, albeit with differences in thickness, lithofacies, and preservation under erosional unconformities (figs. 6, 7). Member contacts define regionally significant sequence boundaries and/or distinct paleoenvironmental shifts, but they are not always distinguishable at a map scale, so we have elected to keep them informal. Below, we describe the primary lithofacies of each member, followed by a brief paleoenvironmental interpretation, and then summarize our findings in a regional tectono-stratigraphic synthesis put forth in the Discussion.

Lowermost strata of the Callison Lake Fm are heterogeneous along depositional strike and in the different inliers (figs. 6, 7). The basal surface is either marked by a significant disconformity with evidence for subaerial exposure, paleokarst development, and intense silicification, or a stratigraphic truncation of underlying units with subtle (<2°) to profound (>30°) angular discordance (fig. 5A-C). In the Coal Creek inlier, the upper contact is either gradational into overlying siliciclastic rocks of the Seela Pass Fm or marked by a profound erosional unconformity with overlying glaciogenic rocks of the Eagle Creek Fm (figs. 4, 6). In the Hart River inlier, the upper Callison Lake Fm is either

truncated by imbricate Mesozoic and younger thrust faults or marked by an erosional
unconformity with various younger units (figs. 3, 7).

Heterolithic member – Description

The Heterolithic mb, named for its diverse lithological composition, is 23.5 to
155.9 m thick. A mixed suite of siliciclastic deposits (Facies F1, F2, F3, F4; table 3)
characterize the bulk of this unit, but it also contains a distinct package of stromatolitic
biostromes and bioherms (F7) that are present in both the Coal Creek and Hart River
inliers (figs. 5H, 6, 7). The base of the Heterolithic mb rests on extensively silicified
carbonate strata of the Craggy dolostone and is commonly marked by a thin (0.05–0.45
m), clast-supported chert and quartz pebble conglomerate unit (F1; fig. 8A). Thick- to
medium-bedded and locally channelized sandstone deposits (F2), interbedded with
variegated shale and siltstone (F3), overlie this coarse-grained unit in a series of stacked
fining-upwards packages that range from ~10–70 cm thick (fig. 8B). The fine- to coarse-
grained sandstone beds are moderately- to well-sorted, consist of quartz or chert arenite
and wacke, and are characterized by crude parallel lamination and sparse symmetrical
ripple- to unidirectional dune-scale trough cross-bedding with abundant shale-chip clasts.
The variegated siltstone and shale facies (F3) contains mudcracks (fig. 8C), planar
lamination, flaser bedding, and rare ball-and-pillow structures. Locally, there are clear
erosional surfaces in the siltstone and shale units capped by crudely normal-graded and
coarser-grained (up to granule size) sandstone beds that eventually fine-upwards back
into mudcracked shale- and siltstone-dominated intervals; these m-scale, coarser-grained
packages tend to have lenticular, channel-like geometries (fig. 5G).

407 The sandstone-dominated interval gradationally transitions into variegated shale
408 and siltstone (F3) interbedded with dolomitic stromatolite biostromes and bioherms (F7),
409 which are capped by a prominent interval of black shale (F4) (figs. 5, 6, 7, 8D-F). The
410 bright yellow-orange to buff-white bioherms and biostromes generally range from ~0.2 to
411 42.8 m thick and are composed of morphologically diverse stromatolites, including
412 conical forms, low inheritance branching structures (fig. 8D), laterally linked and/or high
413 relief domal structures (fig. 8F), and high inheritance digitate forms, all of which have
414 silty dolomudstone intercolumnar fill and local synoptic relief. The doloboundstone units
415 also contain rare frosted quartz grains, and petrographic examination of stromatolite
416 laminae yields zones of disseminated organic matter with dolomitized cyanobacterial
417 sheaths (fig. 8G) and VSMs (fig. 8H). Notably, the Mount Harper East section (J1301)
418 has a much thicker (~40 m thick) stromatolitic buildup than the rest of the sections (figs.
419 5H, 6) – this composite structure appears to comprise multiple superimposed
420 stromatolitic bioherms and could be classified as a reef (*sensu* Geldsetzer and others,
421 1988).

422 The overlying black shale interval commonly drapes the distinct stromatolitic
423 horizons and is up to ~60 m thick. A VSM-bearing shale layer at the base of this unit in
424 the Mount Harper East section (J1301) has been dated with ORR Re-Os geochronology
425 at 752.7 ± 5.5 Ma (fig. 6; Rooney and others, 2015). These strata are characterized by
426 laminated, organic-rich black shale interbedded with thin intervals of gray-green silicified
427 siltstone, planar laminated and silt-rich yellow-orange dolomudstone, and altered zones
428 with abundant hematite, pyrite, sphalerite, and ankerite (figs. 6, 7). The black shale
429 interval in the Coal Creek inlier also contains a thin (~2–25 cm), matrix-supported

430 conglomerate bed with mm- to cm-scale poorly sorted and angular clasts of olive-green
431 shale, chert-replaced isopachous and botryoidal cements, unusual chert spherules, and
432 abundant cm-scale cubic pyrite. In the Hart River inlier, almost all of the fine-grained
433 siliciclastic strata of the Heterolithic mb contain a prominent penetrative cleavage and are
434 difficult to distinguish from underlying deposits of the Fifteenmile Gp equivalents (for
435 example, map unit PPD3; fig. 3); however, this member is consistently present in the two
436 inliers, albeit with prominent thickness and facies variation (figs. 6, 7). The contact with
437 the overlying Talc mb is commonly marked by an abrupt transition into light gray
438 microbial dolostone and/or talc-rich black shale.

440 *Heterolithic member – Interpretation*

441 Mixed siliciclastic and carbonate deposits of the Heterolithic mb record a
442 decameter-scale fining-upwards package from the basal conglomeratic unit into thick
443 black shale deposits. The base of the Heterolithic mb most likely records marine
444 transgression over pre-existing irregular paleokarst topography formed in the underlying
445 Craggy dolostone; in particular, the basal conglomeratic unit is interpreted here to
446 represent a transgressive lag associated with regional deepening in the Callison Lake
447 basin(s). As highlighted below, we tentatively interpret the Heterolithic mb to record
448 sedimentation in a tidal flat to lagoonal depositional setting characterized by low
449 topographic relief; however, the regional poor exposure of the Heterolithic mb, as well as
450 the lack of paleocurrent data and poor preservation of internal bedding structures in
451 sandstone units, precludes a very detailed paleoenvironmental interpretation.

452 The general fine grain size of basal Heterolithic mb deposits reflects predominant
453 deposition from suspension; however, this low-energy depositional regime was clearly
454 disrupted by intervals of higher-energy, bed-load sedimentation characterized by
455 channelized, medium- to fine-grained sandstone units with basal erosional scours and
456 abundant shale-chip conglomerate. The abundance of mudcracks in the variegated shale
457 and siltstone facies (F3), and the presence of discrete shale partings with mudcracks
458 between planar laminated sandstone beds, suggests these fine-grained deposits were
459 commonly exposed to subaerial conditions. This evidence for periodic desiccation, in
460 combination with the presence of local flaser bedding and symmetrical ripple cross-
461 lamination in siltstone and sandstone units, provides support for a potential tidal
462 influence on basal Heterolithic mb strata. Furthermore, the m-scale, crudely cross-bedded,
463 and channelized sandstone units potentially represent intertidal dunes, or tidal bars (for
464 example, Allen and Homewood, 1984; Ashley, 1990), which is supported by the the
465 lateral transition of these coarser-grained units into the episodically-exposed variegated
466 shale and siltstone facies (F3; fig. 5G). These combined features could also reflect low-
467 energy fluvial sedimentation on a coastal plain; however, the fine-grained nature of the
468 deposits, the facies stacking pattern of the entire Heterolithic mb, and the presence of a
469 basal conglomeratic lag all lend greater support to the development of a distinct
470 transgressive, tide-dominated marine succession (Cattaneo and Steel, 2003; Dalrymple,
471 2010; Desjardins and others, 2012 and references therein). Unfortunately, the lack of
472 paleocurrent data precludes the identification of bimodal versus unidirectional current
473 indicators, as well as the development of a precise geometrical configuration for the
474 paleoshoreline.

475 The upsection loss of coarser-grained siliciclastic deposits suggests a gradational
476 transition into a zone of consistent suspension deposition, either due to relative deepening
477 in the basin(s) below fair weather wave base (FWWB) or through the development of an
478 outboard protective barrier to wave action (that is, shoal complex, sand ridge, or barrier
479 island complex). The lack of current-generated bedforms in the upper Heterolithic mb
480 and the development of laterally extensive, thick organic-rich black shale deposits is
481 suggestive of a lagoonal or shelf interior depositional environment. This change in
482 bathymetry is also consistent with the development of high relief and ornate stromatolitic
483 bioherms and biostromes with minimal evidence for reworking through current or wave
484 action. The bounding lithofacies composition and facies architecture in both the basal
485 Heterolithic and overlying Talc members, in combination with the presence of VSMs in
486 both stromatolitic laminae and encasing black shale units (figs. 6, 7, 8H), also implies a
487 marine depositional setting for the upper Heterolithic mb (for example, Porter and Knoll,
488 2000).

489 490 *Talc member – Description*

491 The Talc mb, named for its unusual mineralogical composition (Tosca and others,
492 2011), ranges from 44.9 to 108.8 m in thickness. These strata are characterized by
493 extreme lateral and vertical facies heterogeneity on a cm- to m-scale (figs. 6, 7, 9);
494 therefore, we have simplified the main talc-rich facies into two categories (table 3) that
495 include strata characterized by mm- to m-thick talc-rich black shale (F5) and intervals
496 containing interbedded talc-rich shale and dolostone (F6). Other volumetrically
497 significant facies that characterize this mb include stromatolitic and microbial dolostone

(F8, 9), intraclast grainstone and wackestone (F10), and oolitic or peloidal dolograinstone (F11) (figs. 6, 7, 9; table 3).

Talc-rich shale strata (F5) are dark-gray to jet-black, have a distinctive vitreous luster (fig. 9A), and locally contain abundant black chert nodules or silicified horizons (mm- to cm-thick). The shale is locally truncated by erosional surfaces, which are generally overlain by channelized bodies of shale-clast, matrix-supported conglomerate (F6; fig. 9B) or oolitic and oncolitic dolograinstone (F11). No crossbedding has been identified in these channelized deposits, but crude sub-parallel to inclined lamination is visible in finer-grained oolitic and intraclast dolograinstone lithologies.

Most dolostone units in the Talc mb are composed of mm- to cm-scale nodular dolomudstone (F6), planar to wavy laminated doloboundstone with morphologically diverse stromatolites (F8, F9; fig. 9C), or oncolitic and oolitic dolograinstone (F11). The stromatolites are characterized by a diverse morphological spectrum from cm-scale digitate and cusped structures to large (>2.45 m thick) domal or conical buildups that have clear synoptic relief (fig. 9C); most stromatolitic laminae are coated with thin drapes of talc-rich shale. These microbially influenced and planar laminated dolostone horizons are commonly interbedded with, or grade laterally into, stromatolite-clast rudstone and wackestone, locally overturned stromatolitic mounds suspended in talc-rich shale (fig. 9D), and ptymatically-folded dolostone laminites (fig. 9E). Other dolostone facies in the Talc mb include lenticular, m-scale deposits of medium- to thick-bedded, trough crossbedded oolitic and pisolitic dolograinstone (F11).

The Talc mb also contains tepees (*sensu* Kendall and Warren, 1987), mudcracks, and minor matrix-supported carbonate breccias (F14) at various stratigraphic levels in the

carbonate-dominated strata (figs. 6, 7, 9). Dolomite pseudomorphs after gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$) are locally present, as evidenced by cm-scale disc-shaped crystals and vertically oriented fibrous textures that resemble selenite, or satin spar. Dolomitic replacement fabrics after anhydrite (CaSO_4) are also common in the Talc mb, including nodular bedding resembling chickenwire texture (fig. 9F), discrete zones of enterolithic folding (*sensu* Butler and others, 1982), and isolated to coalesced displacive growth structures (fig. 9F). No primary evaporite minerals were recognized in the Talc mb; however, chert-replaced nodular structures after anhydrite with μm -scale relict anhydrite clusters, zebraic and length-slow chalcedony in megaquartz (Milliken, 1979; Folk and Pittman, 1971; Ulmer-Scholle and Scholl, 1994; Ulmer-Scholle and others, 1993), and small ($< 200 \mu\text{m}$) gypsum laths suspended in organic matter are present in thin section.

The Talc mb facies belt is traceable, albeit discontinuous, between the Coal Creek and Hart River inliers ($>150 \text{ km}$) with discrete intervals characterized by subtle cyclicity and vertically stacked facies associations (figs. 6, 7, 9). For example, $\sim 0.70\text{--}1.45 \text{ m}$ thick parasequences composed of talc-rich shale, stromatolitic doloboundstone, microbialite, and mudcrack-dominated talc-rich shale and nodular carbonate are present in almost every section (figs. 6, 7, 9). Recrystallized sucrosic dolostone and pervasive silicification in the form of chert nodules (mm- to cm-scale) and stratiform chert horizons are also common features of dolomitic strata in the Talc mb.

Talc member – Interpretation

As highlighted below, the Talc mb likely records deposition in a stable, episodically restricted marginal marine setting. The presence of evaporite pseudomorphs,

544 tepees, mudcracks, and the intimate association of microbial dolostone and organic-rich
545 fine-grained strata all suggest deposition in a subaerially exposed, but periodically
546 flooded, peritidal mud flat, or sabkha, similar to the modern Trucial Coast of the Arabian
547 Gulf (for example, Purser, 1973, Bathurst, 1975; Schreiber and others, 1986).

548 The predominance of laminated, talc-rich shale deposits, the lack of distinct
549 wave-generated bedforms, and the abundance of microbial dolostone suggests consistent
550 low-energy suspension and precipitation-based deposition in subtidal to supratidal
551 depositional environments. The isopachous “stromatolitic” laminites of the Talc mb
552 resemble subtidal evaporitic textures previously described by Pope and others (2000) in
553 other Precambrian carbonate-evaporite sedimentary successions. There is also evidence
554 for episodic high-energy events in the Talc mb that are responsible for channel incision,
555 oolitic to oncolitic grainstone deposition, the deformation of microbial and/or suspension-
556 load laminites, and the development of shale-chip conglomerate and
557 stromatolitic/intraclast rudstone. The m-scale, channelized oolitic grainstones (F11) and
558 shale-chip conglomerates (F6) most likely represent tidal channel avulsion and migration,
559 whereas the presence of stromatolitic intraclast grainstone and wackestone and local
560 synsedimentary deformation is more suggestive of episodic storm or seismic activity.

561 A marine paleoenvironmental setting characterized by alternation of subtidal and
562 supratidal deposition is also supported by the abundance of microbial dolostone fabrics,
563 from supratidal, planar-laminated microbialite to subtidal, m-scale stromatolitic buildups
564 with synoptic relief. The distinct parasequence architecture of the Talc mb likely reflects
565 marine shoaling cycles from subtidal talc-rich shale deposits into supratidal microbialite
566 with subaerial exposure surfaces (fig. 9). This parasequence architecture is similar to

those described from examples of modern and ancient marine peritidal deposits (for example, Purser, 1973; Grotzinger, 1989; Grotzinger and James, 2000 and references therein). Finally, the stratigraphic architecture of subjacent Heterolithic and Ramp mb strata also supports a model of tidal flat progradation, which is also characteristic of modern and ancient sabkha depositional reconstructions (for example, Pratt, 2010 and references therein).

The most distinctive component of the Talc mb is its namesake – the widespread presence of talc as an authigenic sedimentary mineral. Tosca and others (2011) performed X-Ray diffraction (XRD) analyses and experiments to show that talc was not only the main constituent of these unusual shale units, but also derived from an early authigenic precipitate due to its distinctive crystallographic disorder. The presence of talc in the Callison Lake Fm is most likely a result of relatively late (burial?) diagenetic transformation from a hydrated Mg-clay precursor such as kerolite ($\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_2 \cdot \text{H}_2\text{O}$), sepiolite ($\text{Mg}_2\text{Si}_6\text{O}_{15}(\text{OH})_2 \cdot 4\text{H}_2\text{O}$), or stevensite ($\text{Na}_{0.15}\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_4$), all of which are metastable relative to talc (Tosca and others, 2011). Most descriptions of modern Al-free Mg-clay occurrences are from evaporitic and alkaline lacustrine basins (Calvo and others, 1999 and references therein); however, they are also described in open marine environments, pedogenic horizons, hydrothermal settings, and weathering profiles of mafic volcanic rocks (Millot, 1970; Weaver and Beck, 1977; Stoessell and Hay, 1978; Yan et al., 2005; Dekov et al., 2008; Galán and Pozo, 2011; Tosca, 2015 and references therein). The significance of the widespread distribution, impressive thickness, and unique depositional fabrics of Callison Lake talc deposits is being presented elsewhere.

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Ramp member – Description

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The Ramp mb is the thickest and most laterally extensive unit in the Callison Lake Fm, measuring up to ~518 m thick near Mount Harper in the Coal Creek inlier (section J1210; fig. 6). The contact between the Talc and Ramp members is marked by the loss of talc-rich shale and a transition into light- to dark-gray thinly bedded dolomudstone with bright red to maroon shale partings (figs. 6, 7, 10B). Exposures of the Ramp mb average ~400 m thick and are overlain by an erosional unconformity (figs. 6, 7), except in localities in the eastern Coal Creek inlier where the overlying Transitional mb is conformable with the Seela Pass Fm (fig. 5).

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Relatively homogeneous light- to dark-gray and medium- to thick-bedded dolostone characterizes the Ramp mb. These dolostones locally appear massive and structureless due to a combination of fabric-destructive dolomitization and surficial caliche precipitation and lichen growth. Characteristic facies include microbial and stromatolitic dolostone (F7-9), oolitic and peloidal dolograinstone (F11), stromatolitic intraclast grainstone and wackestone (F10), laminated dolomudstone and dolosiltite (F12), and diagenetic facies such as pervasively recrystallized dolostone (F13) and carbonate breccia (F14). Similar to microbial structures described in the underlying strata, Ramp mb microbial dolostone is characterized by a pronounced diversity in morphology and size, including flat, crinkly laminated microbialite with fenestral fabrics, laterally-linked to high-relief domal structures (fig. 10E), high-inheritance columnar forms, and m-scale broad domal buildups (fig. 10F).

612 One of the most common features in the Ramp mb is the abundance of
613 stromatolite clast grainstone, wackestone, and floatstone (F10; figs. 6, 7, 10E); these cm-
614 to m-scale crudely laminated and poorly sorted deposits are commonly associated with
615 erosional surfaces in underlying stromatolite or microbialite units (for example, fig 10E).
616 Other possible microbial features in the Ramp mb include cm- to m-thick intervals of
617 mm- and cm-scale “clotted” fabrics with irregularly shaped cavities filled by multiple
618 generations of dolomitic cement (fig. 10C). These unusual “clotted” fabrics resemble
619 features described as “thrombolitic” by Harwood and Sumner (2011; 2012) from the
620 Neoproterozoic Beck Spring Dolomite of Death Valley and putative microbial fabrics
621 mentioned by Turner and others (2011) in the correlative Coppercap Fm of the
622 Mackenzie Mountains, NWT (fig. 2); however, the origin of these structures remains
623 ambiguous in the Callison Lake Fm because we lack clear petrographic evidence for a
624 non-diagenetic origin. The Ramp mb also contains multiple, m-thick, massive oncolitic
625 dolograinstones and floatstones that locally contain pendant cements (fig. 10D).

626 The Ramp mb contains medium- to thick-bedded, fine- to coarse-grained, and
627 trough and tabular cross-bedded dolograinstone horizons (F11; figs. 6, 7, 10A). Giant
628 ooids (Swett and Knoll, 1989; Grotzinger and James, 2000), peloids, pisoids, and various
629 intraclasts are the predominant clasts in these grainstone deposits. The geometries of
630 these deposits are difficult to reconstruct on single ridgeline exposures in the Ogilvie
631 Mountains, but in some instances, one can document distinct along-strike transitions into
632 finer-grained deposits of dolosiltite, dolomudstone, or stromatolitic doloboundstone (figs.
633 6, 7). As noted above, the dolosiltite and dolomudstone units (F12) are generally thin-
634 bedded and contain ferruginous clay partings and local erosional scours (fig. 10B); no

distinct grading has been recognized in these intervals. These finer-grained deposits, as well as the microbialite units (F9), locally display irregular ptigmatic folds and convolute bedding (for example, Mount Gibben East, section J1302; figs. 6, 7, 10G).

Grainstone deposits of the Ramp mb (F10, 11) tend to be the locus of late-stage, fabric-destructive diagenetic recrystallization (F13), most likely due to their elevated primary porosity. These sucrosic dolostones are generally thick-bedded or massive, and commonly devoid of sedimentary structures in outcrop (figs. 6, 7). Some of the massive dolostone units underlie massive carbonate breccias (F14, figs. 6, 7, 10H). The buff-colored, matrix-supported breccias range from ~0.7 to 4.2 m thick and are generally composed of a fitted fabric of angular dolostone clasts (locally silicified) in a matrix of terrigenous silt, dolosiltite, or pure coarse-grained dolomite spar (fig. 10H). Finally, black to dark gray chert is another ubiquitous diagenetic feature in carbonate deposits of the Ramp mb.

Ramp member – Interpretation

As described below, the Ramp mb lithofacies record a mixture of subtidal to supratidal depositional environments. Thick intervals of chaotic, ptigmatically-folded and laminated dolosiltite and microbialite (for example, Mount Gibben East, section J1302; figs. 6, 10G) suggest that Ramp mb strata were subject to episodic synsedimentary deformation and seismic activity. Concordant and planar-laminated strata bound these convoluted intervals, so the deformation was not post-depositional and younger in origin.

The basal fine-grained strata of the Ramp mb most likely reflect sub-wave base suspension deposits; the discrete ferruginous clay laminae could record episodic pulses of

very-fine-grained siliciclastic delivery into the basin and/or the development of basinal hardgrounds and authigenic iron silicate precipitation. Stratigraphically higher intervals of dolomudstone and dolosiltite deposits interfinger with doloboundstone and oolitic dolograinstone units, which is suggestive of deposition in the lee of topographically elevated subtidal features. In support of this interpretation, several components of the Ramp mb record evidence for peritidal deposition and episodic subaerial exposure. For example, the abundance of planar, crinkly laminated microbialite with abundant fenestral fabrics is a characteristic feature of peritidal carbonate deposits (for example, Pratt, 2010 and references therein). These supratidal features, and their intimate association with interbedded dolograinstone and doloboundstone facies, suggests that these represent small emergent tidal flats that formed in the lee of topographic barrier complexes such as oolitic shoals or stromatolitic build ups similar to the Trucial Coast (for example, Purser, 1973).

The abundance of stromatolite intraclast grainstone in the majority of Ramp mb deposits, as well as coarse-grained grainstone of variable allochem composition, is suggestive of a generally high-energy depositional setting above storm wave base (SWB). This is confirmed by distinct erosional scours in doloboundstone facies (for example, fig. 10E), as well as trough cross-bedded oolitic grainstone deposits that require relatively constant subtidal wave action; however, the preservation of poorly sorted intraclast dolowackestone and ornate stromatolitic and microbialite textures between erosional surfaces requires episodes of stromatolite proliferation and lower-energy suspension deposition. This is also supported by the widespread development of m-scale, high relief domal stromatolitic structures in close association with wackestone deposits (figs. 6, 7,

10F). Therefore, these somewhat disparate facies associations most likely reflect deposition at different water depths on a carbonate ramp with the localized development of patch reefs and microbial buildups (*sensu* Read, 1982).

The massive carbonate breccia units in the Ramp mb are interpreted as paleokarst horizons developed during subaerial exposure episodes of the carbonate depositional system. Similar to the peritidal carbonate fabrics, most of these horizons are relatively thin and difficult to trace along strike between individual sections; however, near Mount Harper there is evidence for an extensive ~60 m thick paleokarst unit that defines the upper boundary of the Ramp mb (section J1210; figs. 4, 6; Mustard and Donaldson, 1990). This interval was previously considered to define a major disconformity between the Fifteenmile and Mount Harper groups (Mustard and Donaldson, 1990; Mustard, 1991; Mustard and Roots, 1997); but, approximately 7.5 km along depositional strike at Mount Harper East (section J1301) there is no karst horizon and the Ramp-Transitional mb contact is conformable (fig. 6), reflecting a diachronous contact between these two members.

Transitional member – Description

Previous workers (for example, Thompson and others, 1994; Mustard, 1991; Mustard and Roots, 1997; Macdonald and others, 2010; 2011) included the Transitional mb in the basal part of the Seela Pass Fm because the upper Ramp mb contact delineates a distinct map-scale lithological boundary; however, we demonstrate here that these strata are not only tied to the Callison Lake Fm, but also record one of the most critical geochemical and tectonic transitions in the Mount Harper Gp.

704 The Transitional mb consists of a relatively thin (19.4 to 43.4 m thick) and
705 discontinuous package of mixed siliciclastic and carbonate rocks that only outcrop in the
706 eastern part of the Coal Creek inlier and define a coarsening upwards trend into maroon
707 shale, sandstone, and conglomerate of the Seela Pass Fm (figs. 4, 6). West of Mount
708 Harper the Ramp-Transitional mb contact is either truncated by overlying younger units
709 or marked by a significant disconformity and paleokarst horizon (figs. 3, 6; Mustard and
710 Donaldson, 1990); in contrast, east of Mount Harper, this contact appears to be
711 conformable, although with some component of stratigraphic onlap as discussed below.
712 In the conformable sections, the member contact is typically marked by an abrupt
713 transition from interbedded oolitic dolograstone and doloboundstone (F8, 9, 11) of the
714 upper Ramp mb into very-thin-bedded dolomudstone and dolosiltite (F12; fig. 6). These
715 fine-grained deposits range from 5–70 cm thick and are characterized by wavy and
716 tabular bed geometries with 0.1–0.9 cm thick maroon clay partings and abundant black
717 chert nodules.

718 The bulk of Transitional mb deposits generally consist of interbedded dark-gray
719 stromatolitic dolostone (F8) and black shale (F4). Stromatolites of this interval are
720 characterized by a variety of digitate columnar forms (fig. 11A), high-inheritance and
721 laterally linked domal structures (fig. 11B), and irregularly shaped branching
722 morphologies; most of these stromatolites are locally affected by zones of convolute
723 lamination, slumped features, or healed syndimentary, cm-scale normal faults (fig.
724 11B). The stromatolites also tend to be in primary growth position, contain significant
725 fine-grained terrigenous silt and clay components, and lack erosional surfaces. These
726 deposits are then overlain by a package of laminated and silicified black shale and

siltstone (F4) that ranges from 9.8 to 11.3 m thick (fig. 6). On an outcrop scale, these poorly exposed and organic-rich siliciclastic deposits appear to be homogeneous and very-fine grained; however, in thin section they are clearly composed of poorly sorted and crudely stratified mud, silt, and very-fine sand-sized particles (fig. 11C) – no evidence for grading has been documented. At Mount Gibben East (section J1204; fig. 6), a black shale layer from this interval yielded a Re-OS age of 739.9 ± 6.1 Ma and contained assemblages of diverse VSMs (figs. 6, 11C; Strauss and others, 2014a).

There are exceptions to this general lithofacies progression. For example, at Mount Harper (J1210) the Ramp-Transitional mb contact is marked by an abrupt transition from massive silicified breccia into interbedded black shale (F4) and very thin- to medium-bedded, extensively fractured orange-yellow laminated dolomudstone and dolosiltite (F12; fig. 6). This interval is both cut by numerous mafic sills of the Mount Harper Volcanics (fig. 6) and sheared between rheologically stiff bounding units (Ramp mb and Seela Pass Fm); however, it appears to be equivalent to the prominent black shale horizon midway through the Transitional mb and is therefore missing the basal carbonate-shale sequence (fig. 6).

The uppermost Transitional mb consists of interbedded microbialite (F9), stromatolite doloboundstone (F8), stromatolite intraclast grainstone and wackestone (F10), and silicified breccias (F14) (fig. 6). Microbial and stromatolitic dolostones are light- to dark-gray and consist of stratiform crinkly lamination and laterally-linked, low relief stromatolitic domal structures up to 35 cm tall. These microbialites are locally truncated and interbedded with thin- to medium-bedded stromatolite intraclast wackestone and grainstone that fill topographic relief above erosional scours. Some of

these microbialite intervals display fenestral fabrics, mudcracks, and tepee structures (fig. 6). At Mount Gibben East (J1203 and J1204), these carbonate strata are capped by enigmatic, ~4-10 m thick silicified dolostone breccia units. The massive matrix-supported breccias are characterized by poorly sorted, pebble- to boulder-sized angular to subrounded clasts of extensively silicified dolostone that resemble facies of the underlying Ramp and Transitional mb strata (fig. 11D). The base of the overlying Seela Pass Fm is defined herein as the first appearance of sandstone or conglomerate, which is laterally variable in stratigraphic location due to the NE-directed progradation of Seela Pass alluvial systems over Transitional mb deposits (for example, Mustard, 1991).

Transitional member – Interpretation

The abrupt shift from upper Ramp mb mixed subtidal and supratidal deposits into finer-grained strata of the basal Transitional mb records a prominent deepening into a zone of low-energy suspension sedimentation. We interpret the ferruginous clay partings to represent a combination of hardground cementation and episodic fine-grained siliciclastic rainout. The overlying interval of interbedded black shale and stromatolitic doloboundstone lacks erosional surfaces, storm deposits, and wave- or current-generated bedforms, which suggests these strata were either deposited below SWB or isolated from wave action through the development of a topographic barrier (for example, shoal complex, sand ridge, or barrier island complex). Onlap of these strata onto the prominent paleokarst horizon at Mount Harper suggests base-level rise was involved in this distinct facies shift, and the development of a protected lagoonal or shelf interior depositional setting is supported by the absence of a transition through distinct shoreface deposits and

the along-strike development of extensive organic-rich black shale deposits. Continuous deepening below SWB and/or increased isolation from wave action coupled with drowning of the carbonate factory or burial due to siliciclastic influx is consistent with the upsection loss in carbonate and deposition of laminated black shale and siltstone. However, this deepening may also be the result of local synsedimentary faulting as suggested by the abundance of siliciclastic material in the stromatolitic units and the presence of cm-scale healed normal faults and convolute lamination, as well as the limited geographic range of this unit.

The upper part of the Transitional mb records shallowing from the underlying suspension-dominated subtidal environments into a zone of supratidal carbonate sedimentation. The reappearance of stromatolitic doloboundstone with erosional scours, intraclast wackestone and grainstone, and fenestral microbialite with sparse mudcracks and tepees is characteristic of subtidal to supratidal depositional settings. The development of subaerial conditions is consistent with the eventual widespread progradation of Seela Pass alluvial fan and fan delta deposits over the Transitional mb. Although one could interpret the silicified breccias at Mount Harper East as paleokarst units, the presence of clasts from the underlying Ramp and Transitional members makes these breccias difficult to explain as *in situ* dissolution breccias. Another possibility that we discuss below is that they represent matrix-supported debrites, consisting of reworked fault talus from localized, basin-bounding structures associated with the onset of regional extension.

SEQUENCE STRATIGRAPHY OF THE CALLISON LAKE FORMATION

Sequence stratigraphy has been applied previously to Proterozoic sedimentary successions in NW Canada to correlate amongst disparate stratigraphic sections in geographically isolated basins (Grotzinger, 1986; Bowring and Grotzinger, 1992; Rainbird, 1993; Long and others, 2008; Macdonald and others, 2012; Thomson and others, 2015a). Here, we provide a basic sequence stratigraphic framework for the Callison Lake Fm based on the recognition and interpretation of transgressive-regressive (T-R) cycles in outcrop exposures from the Coal Creek and Hart River inliers. T-R cycles (that is, depositional sequences) are comprised of discrete packages of strata deposited during a full cycle of change in accommodation or sediment supply (*sensu* Catuneanu and others, 2009; 2011). These cycles are bounded by subaerial unconformities, flooding surfaces, shoreline ravinement surfaces, or regressive marine erosional surfaces, and they contain a transgressive phase that records an upward deepening event and a regressive phase that records an upward shallowing event (for example, Johnson and Murphy, 1984; Johnston and others, 1985; Embry and Johannessen, 1992; Embry, 1995; 2009). Here, we follow the terminology put forth by Catuneanu and others (2011) to define subaerial unconformities, maximum flooding surfaces, maximum regressive surfaces, and transgressive ravinement surfaces.

The Callison Lake Fm records three discrete T-R cycles (fig. 12; labeled T-R6, T-R7, and T-R8). This terminology builds upon previous sequence stratigraphic work in NW Canada from Long and others (2008) and Thomson and others (2015) in the correlative upper Shaler Supergroup of the Amundsen Basin, NWT. The basal Heterolithic mb deepening over the Craggy dolostone-Callison Lake subaerial unconformity records the onset of T-R6. We interpret the conglomeratic lag at the base of

the Heterolithic mb to represent a transgressive ravinement surface associated with T-R6, and the maximum flooding surface (MFS) is recorded in black shale deposits just above the prominent stromatolitic bioherms and biostromes of the middle Heterolithic mb (figs. 6, 7, 12). The overlying Talc mb belongs to the regressive part of T-R6 and records a distinct episode of tidal flat progradation and parasequence development (fig. 12).

The base of T-R7 is marked by a sharp transition from evaporitic dolostone and talc-rich shale of the Talc mb into non-restricted carbonate deposits of the Ramp mb (fig. 12). This boundary most likely represents a maximum regressive surface (*sensu* Embry, 2009), as it records the onset of marine transgression over evaporitic strata of the Talc mb. A thin interval of thin-bedded, laminated dolomudstone and dolosiltite at the base of the Ramp mb represent the transgressive phase of T-R7 (figs. 6, 7, 12). The MFS of T-R7 is difficult to place, but the regressive phase of T-R7 is clearly defined by the thick and extensive progradational carbonate strata of the Ramp mb (fig. 12). A local subaerial unconformity defines the Ramp-Transitional mb contact and marks the top of T-R7 (figs. 6, 12).

T-R8 is complicated by local extensional tectonism during the Callison Lake-Seela Pass transition. Transgressive deposits associated with T-R8 include thin-bedded dolomudstone and interbedded microbial dolostone and black shale of the basal Transitional mb (fig. 12). The MFS of T-R8 is somewhere in the prominent black shale deposits of the middle Transitional mb, which locally drape the subaerial unconformity surface in the western part of the Coal Creek inlier (fig. 6). Regressive deposits of T-R8 are represented by the transition back into microbial carbonate deposition at the top of the

Transitional mb; the upper T-R8 boundary is most likely represented by a local subaerial unconformity at the Callison Lake-Seela Pass formational boundary (fig. 12).

CARBONATE CARBON AND OXYGEN ISOTOPE RESULTS

Previously published $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ chemostratigraphic data for the Callison Lake Fm are reported from two stratigraphic sections in the Coal Creek inlier and define a trend from relatively enriched background $\delta^{13}\text{C}_{\text{carb}}$ values ($\sim 5\text{‰}$) to a negative anomaly down to approximately -6‰ (Mount Harper and Mount Gibben East, fig. 4; Macdonald and others, 2010; Strauss and others, 2014a). Here, we present 1222 new $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ measurements from the Coal Creek and Hart River inliers that range from -5.8 to 6.1‰ and -11.4 to 4.0‰ , respectively, and are shown in stratigraphic context in figures 6 and 7. Figure 13 provides a compilation of $\delta^{13}\text{C}_{\text{carb}}$ vs. $\delta^{18}\text{O}_{\text{carb}}$ crossplots for the different members of the Callison Lake Fm.

Carbon and oxygen isotope data from the Heterolithic mb range from -1.6 to 5.4‰ and -11.4 to 3.8‰ , respectively (figs. 6, 7, 13). The majority of $\delta^{13}\text{C}_{\text{carb}}$ values are moderately enriched (average = 2.2‰) and almost exclusively come from the stromatolitic biohermal and biostromal intervals in the central portion of the Heterolithic mb; few data come from isolated thin-bedded dolomudstone interbedded with thicker siliciclastic-dominated intervals, most of which tend to be slightly depleted in $\delta^{13}\text{C}_{\text{carb}}$. Oxygen isotope data from the Heterolithic mb, and the Callison Lake Fm in general, are very enriched compared to average Neoproterozoic limestone data and normal to slightly enriched in comparison to average dolostones (for example, Jacobsen and Kaufman, 1999; Jaffrés and others, 2007; Prokoph and others, 2008). Contrary to all the other

members in the Callison Lake Fm, $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ weakly covary in the Heterolithic mb (fig. 13).

Talc mb $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ data are more variable with values ranging from -5.6 to 5.5‰ and -7.1 to 3.8‰, respectively, and do not covary. The majority of Talc mb strata commence with moderately enriched $\delta^{13}\text{C}_{\text{carb}}$ values and trend towards more depleted isotopic compositions upsection (figs. 6, 7). In the Hart River inlier, there is a distinct negative $\delta^{13}\text{C}_{\text{carb}}$ isotope anomaly in the upper Talc mb down to values as low as -5.61‰; this trend is evident in the Coal Creek inlier as well, although it is not as clearly developed. Some stratigraphic sections appear to record a series of $\delta^{13}\text{C}_{\text{carb}}$ shifts in the Talc mb (for example, section J1302, Mine Camp, fig. 6), portions of which could be captured in other sections given the profound thickness differences and sampling resolution.

Ramp mb $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ values range from -1.8 to 6.1‰ and -8.2 to 5.5‰, respectively. $\delta^{13}\text{C}_{\text{carb}}$ data generally record a reproducible positive-drifting trend just above the Talc-Ramp mb transition and remain between 2–5‰ for the entirety of the Ramp mb (figs. 6, 7). However, the middle portion of the Ramp mb in the Hart River inlier displays a distinct interval of stratigraphically consistent depleted $\delta^{13}\text{C}_{\text{carb}}$ values that range from 0 to -1.8‰; this feature is not seen in correlative strata of the Coal Creek inlier. $\delta^{18}\text{O}_{\text{carb}}$ values from the Ramp mb are variable, but also are enriched (ave. = -0.2‰) with $\delta^{13}\text{C}_{\text{carb}}$ vs. $\delta^{18}\text{O}_{\text{carb}}$ data displaying a lack of covariance (fig. 13).

Carbon and oxygen isotope data from the Transitional mb range from -5.8 to 4.2‰ and -9.0 to 4.0‰, respectively. Carbon isotope values from these strata record a prominent ~10‰ negative carbon isotopic excursion that has been previously correlated

with the global ca. 735 Ma Islay anomaly (Prave and others, 2009; Macdonald and others, 2010; Rooney and others, 2014; Strauss and others, 2014a). Here, we report $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ data from four parallel sections through this anomaly that reproduce a prominent trend from enriched $\delta^{13}\text{C}_{\text{carb}}$ values of the upper Ramp mb into variably depleted $\delta^{13}\text{C}_{\text{carb}}$ isotopic compositions in the middle Transitional mb (fig. 6). All of these sections display distinct structure to the negative $\delta^{13}\text{C}_{\text{carb}}$ anomaly, in which the nadir of the excursion is generally associated with m-scale variation between approximately -2 and -5‰ and then a distinct return to enriched values up to 0.8‰ in the uppermost Transitional mb strata. Carbon and oxygen isotope data from the Transitional mb do not covary (fig. 13), and $\delta^{18}\text{O}_{\text{carb}}$ values are similar to those reported in underlying members of the Callison Lake Fm.

DISCUSSION

Tectono-stratigraphic Model for the Callison Lake Formation

The development of a regional angular unconformity beneath the Mount Harper Gp, coupled with previous documentation of fault-controlled sedimentation in the Seela Pass Fm, provides evidence for mid-Neoproterozoic (late Tonian) extensional tectonism in Yukon (Mustard, 1991; Abbott, 1997; Mustard and Roots, 1997; Thorkelson and others, 2005; Macdonald and Roots, 2010; Macdonald and others, 2012). Previous studies placed the onset of extension in the Seela Pass Fm and overlying Mount Harper volcanics of the Mount Harper Gp (for example, Mustard, 1991; Mustard and Roots, 1997), but we suggest the Callison Lake Fm represents an earlier manifestation of regional extension in NW Canada associated with the onset of Windermere Supergroup

910 sedimentation. Therefore, the 752.7 ± 5.5 Ma Re-Os depositional age from the maximum
911 flooding interval of the Heterolithic mb (fig. 12; Rooney and others, 2015) provides an
912 important new temporal constraint for an episode of regional extension in Yukon. Below,
913 we use the sedimentological data described above to reconstruct a depositional model for
914 the Callison Lake Fm (fig. 14); this is followed by a review of regional map patterns and
915 correlations to help refine our kinematic model for mid-Neoproterozoic extensional
916 tectonism in NW Canada and the greater North American Cordillera.

917 The precise geometry of the original Callison Lake basin is difficult to reconstruct
918 due to limited exposure and subsequent erosional truncation and fault reactivation;
919 however, many of the sedimentological features and map patterns point to deposition in
920 semiarid, marine-influenced extensional sub-basins. Evidence for fault related
921 sedimentation includes lateral and longitudinal basin-fill asymmetry, abrupt facies and
922 thickness change, sedimentary structures interpreted as indicating synsedimentary
923 deformation, and the development of localized disconformities and topographic relief.
924 The general increase in member thickness from NNW to SSE in both the Coal Creek and
925 Hart River inliers suggests regional deepening to the SSE towards hypothesized ancestral
926 basin-bounding structures (figs. 6, 7, 14). The lack of exposure between the Hart River
927 and Coal Creek inliers makes it difficult to assess whether these two areas were originally
928 linked by a single, >110 km long extensional basin or if they formed as independent half
929 graben separated by an accommodation, or transfer zone. Given the common
930 segmentation of large, active fault zones (for example, Jackson and White, 1989), we
931 consider these two regions as representing discontinuous marine embayments, or gulfs,

932 formed in hangingwall depocenters and associated with discrete, ~20-40 km-long basin-
933 bounding fault segments of an evolving border fault zone.

934 The general fine-grained nature of Heterolithic mb strata and the lack of locally
935 preserved alluvial fan deposits suggests that the preserved portions of the Callison Lake
936 hangingwall depocenters were relatively distal to basin-bounding structures and ancestral
937 topographic highlands (fig. 14A). Heterolithic mb siliciclastic strata record deposition in
938 a tidally influenced nearshore or coastal plain setting whose sediment supply was
939 potentially sourced by axial fluvial systems and/or hangingwall fans (fig. 14A). The
940 transition from these marginal marine deposits into the upper Heterolithic mb black shale
941 and stromatolitic bioherm facies records regional deepening and the development of a
942 protected lagoonal or interior shelf depositional environment. Shallower, intertidal
943 lithofacies are recorded from sections in the NW Coal Creek inlier (for example, Mine
944 Camp (J1302) and Talc Falls (F926-8); figs. 6, 9), which could represent the local
945 transition into nearshore, supratidal depositional environments. Interestingly, Carr and
946 others (2003) describe very similar fine-grained lithofacies in the early Miocene Nukhul
947 Fm of the Egyptian Suez Rift (Red Sea), which consists of estuary mouth sandstone
948 facies interbedded with estuary funnel and deltaic variegated mudstone and sandstone
949 facies deposited in narrow hangingwall depocenters. Although there is no evidence for
950 the development of such narrow basins in the Callison Lake Fm, we argue that this
951 paleoenvironmental setting potentially provides a good analog for early Heterolithic mb
952 sedimentation.

953 The abrupt paleoenvironmental shift from suspension-dominated deposits of the
954 upper Heterolithic mb into peritidal deposits of the Talc mb most likely represents

regional shoaling and subsequent tidal flat progradation over the shelf/lagoonal depositional system, similar to the Holocene development of the Abu Dhabi sabkhas (for example, Purser, 1973). There are a number of important characteristics of the Talc mb that indicate a marine depositional setting despite some of the ambiguity derived from the presence of abundant authigenic Mg-clay minerals (for example, Calvo and others, 1999). First, the Talc mb facies belt is unlike classic bulls-eye evaporite patterns developed in continental settings (for example, Warren, 1989). Furthermore, there is no evidence for the precipitation of bittern salts or saline carbonates in the Talc mb and the close association of organic-rich, microbially influenced dolostone with sulfate evaporite replacement fabrics is a classic feature of marine-fed sabkhas (for example, Schreiber and El Tabakh, 2000 and references therein). For example, the abundance of dolomite-replaced evaporite pseudomorphs, chaotic displacement fabrics, and m-scale parasequences are all indicative features of intra-sediment sulfate evaporite precipitation and the development of evaporitic cycles, which are ubiquitous features of modern and ancient marine sabkhas (for example, Shearman, 1978; Butler and others, 1982; Kirkham, 1997). Building upon the analog depositional setting of African-Arabian extension in the Red Sea, the Talc mb shares many similarities with early Miocene discontinuous sulfate evaporite deposits of the Gulf of Suez and NW Red Sea, which were deposited in marine-influenced hangingwall depocenters during the earliest phases of Oligo-Miocene extension (Orszag-Sperber and others, 1998 and references therein).

The abrupt transition from evaporitic strata of the Talc mb into laminated dolomudstone of the basal Ramp mb records a prominent flooding event that initiates non-restricted carbonate sedimentation in the Callison Lake Fm (fig. 14B). This base-

level transgression was most likely driven by near- or far-field extensional tectonism, which may be locally evidenced by a concentration of seismically disrupted strata near the top of the Talc mb in certain sections (figs. 6, 7). Despite the erosional truncation of the Ramp mb in the Coal Creek inlier, one can still reconstruct a facies progression of NNW-to-SSE deepening from thinner peritidal-dominated sections (Mine Camp (J1302), fig. 6) to extensive subtidal-dominated sections (Mount Harper East (J1301), fig. 6). This depositional pattern appears to reflect the local development of a hangingwall carbonate ramp (fig. 14B; Read, 1982; Leeder and Gawthorpe, 1987; Bosence, 1998); however, we cannot rule out the possibility that this pattern instead reflects the development of local subaqueous topography in an extensive, rimmed platform-type setting (for example, Read, 1982; Tucker, 1985). Unfortunately, the key deposits to assess these different facies models would have been situated closer to the eroded (or displaced) ancestral basin-bounding fault (fig. 14), where one would expect to either encounter distal starved basin deposits (ramp) or proximal nearshore strata (rimmed platform). Depositional models consistent with the Red Sea analog include either the development of small carbonate depocenters in blind-headed gulfs, similar to upper portions of the Early Miocene Nukhul and Tayran formations of the Gulf of Suez and NE Red Sea (Hughes and others, 1992; Montenat and others, 1998; Bosworth and other, 2005 and references therein), or a temporary divergence to broad rift subsidence recorded in the Rudeis, Burqan, and Habab formations and lower Maghersum Gp of the Red Sea region (Bosworth and others, 2005 and references therein).

The Ramp-Transitional mb boundary represents the onset of tectonic reorganization in the Callison Lake basin(s) that eventually evolved into widespread

extensional faulting and volcanism associated with the Seela Pass Fm and Mount Harper volcanics (fig. 14C). The increase in subaerial exposure surface frequency, the development of local topographic highs, and the widespread evidence for soft-sediment deformation in the upper Ramp mb (figs. 6, 7) provides sedimentological evidence for regional emergence and disruption of carbonate ramp sedimentation. In the Coal Creek inlier, the onset of this synsedimentary tectonism is marked by the coeval development of subaerial exposure and karst in the western uppermost Ramp mb deposits and apparently synchronous rapid deposition of basal Transitional mb strata in the eastern portion of the inlier (fig. 6). These divergent depositional histories indicate rotational motion of the ancestral basin-bonding structure (and its hangingwall depocenter) about a vertical W-E or NW-SE axis. Syntectonic Transitional mb strata were deposited in an evolving and complex paleoenvironmental setting possibly characterized by a narrow, marine gulf fed by antecedent fluvial systems that were eventually buried by alluvial fan and fan delta deposits of the Seela Pass Fm (fig. 14C). The maximum flooding interval of the Transitional mb black shale deposits has been dated with Re-Os geochronology at 739.9 ± 6.1 Ma (Strauss and others, 2014a), which provides an important age constraint for the timing of syntectonic sedimentation close to the Callison Lake-Seela Pass contact.

Fault-associated rotation of the Coal Creek hangingwall depocenter could have been driven by a number of local extensional processes, including uplift at the western fault tip or segment boundary, the regional development or activation of accommodation zone faulting in the western Coal Creek inlier associated with some component of oblique extension, and/or uplift and doming associated with the emplacement of the Mount Harper volcanics at depth (fig. 14C). This phase of syn-Callison Lake tectonism

was possibly presaged by motion along blind structures at depth – the thinning of Ramp
mb carbonate strata towards the NW could represent the development of a hangingwall
monocline above a blind fault and the thickening of strata to the SE could indicate the
coeval growth of a hangingwall syncline (for example, Gawthorpe and Leeder, 2000;
Sharp and others, 2000). Interestingly, regional map patterns highlight a distinct synform
in pre-Seela Pass strata in the Coal Creek subsurface – this previously buried structure
possibly daylights in Seela Pass time as the “Trap Door” fault (fig. 4; Strauss and others,
2014b). Post ~740 Ma, northward-directed progression of the ancestral basin bounding
structures and coeval segmentation of the Callison Lake depocenter(s) is marked by the
development of the Harper Fault and Seela Pass syn-rift deposits in the Coal Creek inlier
(figs. 4, 14C; Mustard, 1991; Mustard and Roots, 1997). This newly developed
extensional half-graben contains its own unique history of fault-related sedimentation,
volcanism, and dike emplacement (Roots, ms, 1987; Mustard, 1991; Mustard and Roots,
1997). Seela Pass equivalent rift-related sedimentation is also present in the Hart River
inlier but has not been documented at the same level of detail (Abbott, 1997). In
conclusion, the Callison Lake basin setting is similar to initial sedimentary deposits in the
Oligo-Miocene northern Red Sea and Gulf of Suez extensional systems, recording no
evidence for widespread doming or significant relief from rift flank uplift, a distinctly
fine-grained siliciclastic syn-rift fill with early carbonate and sulfate evaporites in wedge-
shaped half-grabens, and no apparent relationship to plume-related rifting.

Regional Correlations of the Windermere Supergroup

The Mount Harper Gp shares a similar history of stratigraphic interpretation with the Coates Lake Gp of the Mackenzie Mountains: both units were also originally placed in the Mackenzie Mountains Supergroup and then relocated to the basal Windermere Supergroup after recognition of their rift-related nature (fig. 2; Young and others, 1979; Aitken, 1981; Jefferson, ms, 1983; Jefferson and Ruelle, 1986; Jefferson and Parrish, 1989; Abbott, 1997; Thorkelson and others, 2005; Macdonald and Roots, 2010; Macdonald and others, 2011; 2012; Turner and others, 2011). Coates Lake strata are subdivided into the Thundercloud, Redstone River, and Coppercap formations, and they are separated from the underlying Little Dal Gp by a significant erosional unconformity and discontinuous outcrops of the Little Dal basalt (figs. 1, 2, 15; Jefferson, ms, 1983). Previously published age constraints for the Little Dal basalt (fig. 2) are through geochemical correlation to a fault-bounded $777.8 \pm 2.5/-1.8$ Ma quartz diorite (U-Pb zircon; Jefferson and Parrish, 1989) and the 779.5 ± 2.3 Ma Tsezotene sills (U-Pb baddeleyite; Harlan and others, 2003); however, Milton and others (2015) recently reported a new U-Pb CA-ID-TIMS zircon age of 774.93 ± 0.54 for these mafic flows. The Little Dal basalt and Tsezotene sills have been correlated with the Hottah sheets of the Wopmay Orogen, suggesting a link to the ca. 780 Ma Gunbarrel LIP (Ootes and others, 2008; Sandeman and others, 2014).

The Coates Lake Gp was deposited in a series of marine embayments formed in extensional half-grabens (Jefferson, ms, 1983). The Thundercloud and Redstone River formations are characterized by abrupt lateral facies change between talc-bearing sulfate evaporites (Abercombie, ms, 1978), conglomerate, sandstone, and thin carbonate units, which are suggestive of deposition in proximal to distal alluvial fans, fan deltas, and

marginal marine to playa lake settings (Ruelle, 1982; Jefferson, ms, 1983; Jefferson and Ruelle, 1986). These heterolithic strata gradationally transition into deep-water carbonate-dominated deposits of the Coppercap Fm (Jefferson, ms, 1983), which has been dated with Re-Os geochronology at 732.2 ± 3.9 Ma (figs. 1, 2, 15; Rooney and others, 2014). Various strata of the ~717–662 Ma Rapitan Gp (Macdonald and others, 2010; Rooney and others, 2014) unconformably overlie Coates Lake Gp strata with local angular discordance, attesting to synsedimentary tectonism throughout Coates Lake and Rapitan time (Eisbacher, 1977; 1981; Helmstaedt and others, 1979; Aitken, 1981; Jefferson, ms, 1983; Jefferson and Parrish, 1989; Turner and others, 2011).

Mount Harper equivalent strata of the Shaler Supergroup in the Minto inlier of Victoria Island include the Kilian and Kuujjua formations (figs. 1, 2, 15; Macdonald and others, 2011; 2012; Thomson and others, 2014), which were deposited in the intracratonic Amundsen Basin (Young, 1981; Rainbird, ms, 1991). The Kilian Fm consists of mixed siliciclastic, evaporite, and carbonate strata that record subtidal to peritidal sedimentation in a carbonate ramp and sabkha depositional setting (Young, 1981; Jefferson, 1985; Rainbird, ms, 1991; Rainbird, 1993; Jones and others, 2010). The overlying Kuujjua Fm is dominated by coarse-grained quartz arenite and minor interbedded fine-grained sandstone, shale, and dolomitic siltstone and represents a profound shift to fluvial sedimentation (Young, 1981; Jefferson, 1985; Rainbird, ms, 1991; 1992). Extensive continental flood basalts of the Natkusiak Fm (Thorsteinsson and Tozer, 1962) unconformably to conformably overlie the Kilian and Kuujjua formations and have been dated with baddeleyite on coeval sills at $723 \pm 4/-2$ Ma (U-Pb TIMS; Heaman and others, 1992) and 716.33 ± 0.54 Ma (CA-ID-TIMS; Macdonald and others,

2010); these mafic volcanics, and their associated gabbroic sills and dikes, represent remnants of the Franklin LIP (Heaman and others, 1992). Both the Kilian and Kuujjua formations display erosional truncation, NE stratigraphic thinning, and local evidence for extensional faulting, which have been attributed to pre-eruptive thermal uplift associated with the emplacement of the Franklin LIP (Rainbird, 1993). A maximum age constraint for the Kilian Fm comes from a Re-Os ORR age of 761 ± 41 Ma in the underlying Wynnai Fm (van Acken and others, 2013) and ca. 800 Ma detrital zircon grains in sandstone from the basal Kilian Fm (figs. 1, 2, 15; Rayner and Rainbird, 2013).

Preliminary sequence stratigraphic correlations among Neoproterozoic strata in NW Canada are still being developed (Long and others, 2008; Macdonald and others, 2012; Thomson and others, 2015a). There is no published sequence stratigraphic data from the Coates Lake Gp and previous work has retained the Kilian and Kuujjua formations in Sequence B of Young and others (1979) (Rainbird, 1993; Long and others 2008; Thomson and others, 2015a). Given the significance of the Mackenzie Mountains-Windermere Supergroup boundary throughout NW Canada, we argue that the Kilian and Kuujjua formations should be stratigraphically placed in Sequence C of Young and others (1979) (see discussion by C.W. Jefferson in Long and others, 2008); therefore, we abandon the sequence stratigraphic nomenclature of Long and others (2008) that retains these units in “sub-sequence sB5 of Sequence B” and develop a preliminary T-R cycle correlation scheme based on the work of Rainbird (1993). We also use the detailed sedimentological and stratigraphic data of Jefferson (ms, 1983) to provide a template for preliminary sequence stratigraphic interpretations in the Coates Lake Gp (fig. 15).

Rainbird (1993) documented four submergent-emergent cycles in the Kilian Fm, the fourth of which was interrupted by an abrupt shift to fluvial sedimentation in the Kuujjua Fm. The scale and style of these cycles, as well as their association with basin-wide subaerial exposure surfaces, suggests similarity to the T-R cycles described herein from the Callison Lake Fm. We posit that the three depositional sequences in the Callison Lake Fm (T-R6, T-R7, and T-R8; fig. 12) are equivalent to the initial three cycles of Rainbird (1993) (fig. 15). This is supported by the presence of the Islay Carbon Isotope Excursion (ICIE) in the upper part of T-R8 in both the Callison Lake and Kilian formations (fig. 15; Jones and others, 2010; Macdonald and others, 2010; Prince, 2014; Strauss and others, 2014a; Thomson and others, 2015b). Interestingly, Eisbacher (1977; 1981) and Jefferson (ms, 1983) independently documented three unconformity-bound depositional cycles in the Coates Lake Gp, each later divided into distinct formation boundaries (Jefferson, ms, 1983). The ICIE, which has already been correlated geochronologically between the Coppercap and Callison Lake formations (Strauss and others, 2014a), is also preserved in the third depositional cycle (T-R8) of the Coates Lake Gp (fig. 15). Although this integrated sequence stratigraphic and chemostratigraphic correlation scheme is speculative, it provides an appealing alternative to simplified lithostratigraphic correlation in these tectonically active basins.

Regional sequence stratigraphic correlations of younger depositional cycles in NW Canada are potentially complicated by the onset of diachronous regional extension associated with the emplacement of the Franklin LIP (fig. 15). The base of cycle T-R9 in the Mount Harper Gp is most likely marked by the transition to fault-related sedimentation in the Seela Pass Fm (figs. 12, 15), whereas cycle T-R9 in the Shaler

Supergroup is quite similar in stratigraphic architecture to underlying T-R cycles in the Kilian Fm (Rainbird, 1993). The abrupt shift from marginal marine to fluvial sedimentation at the Kilian-Kuujua contact most likely represents a younger depositional cycle boundary (that is, base of T-R10) and could be similarly driven by the onset of regional extension-related faulting in the Shaler Supergroup (Rainbird, 1993; Prince, 2014; Thomson and others, 2015a). This may be indicative of diachronous, NE-propagating mid-Neoproterozoic extension across NW Canada during the break-up of Rodinia, culminating with the emplacement of the Franklin LIP (for example, Rainbird and others, 2014). Sequence T-R9 does not appear to be present in the Coates Lake Gp due to erosional truncation beneath the Rapitan Gp (fig. 15; Jefferson, ms, 1983). Despite previous studies that suggest a eustatic origin for these T-R cycles (Rainbird, 1993; Long and others, 2008; Thomson and others, 2015a), the regional correlation of these sequence boundaries in NW Canada and their clear association with extension-related unconformities is perhaps more suggestive of a greater tectonic driving mechanism.

Application of this tectono-stratigraphic correlation scheme to other mid-Neoproterozoic strata of the Windermere Supergroup in Laurentia, such as the ChUMP basins of the western U.S., enables a margin-wide comparison. A preliminary sequence stratigraphic scheme for the Uinta Mountain Gp was proposed by Dehler and others (2010) and expanded upon by Kingsbury-Stewart and others (2013), both of which concluded that Uinta Mountain Gp (and correlative Big Cottonwood Fm) siliciclastic strata record three, km-scale fining-upwards depositional sequences (*composite* sequences of Kingsbury-Stewart and others, 2013). A preliminary four-fold, km-scale sequence stratigraphic framework was also proposed for the Chuar Gp (Dehler and others,

2001), although the recognition of ca. 780 Ma detrital zircons in the Nankoweap Fm requires an update to the sequence stratigraphic architecture of these strata (Dehler and others, 2012). No sequence stratigraphy has been reported from the Pahrump Gp of Death Valley; however, recent work by Macdonald and others (2013b), Mahon and others (2014), and Smith and others (2015) has documented an equivalent basal Windermere Supergroup tectono-stratigraphic package that comprises the Horse Thief Springs Fm, Beck Spring Dolomite, and unit KP1 of the Kingston Peak Fm (tectonostratigraphic unit two (TU2) of Macdonald and others, 2013b). The presence of the ICIE in the uppermost Beck Spring Dolomite (Horodyski and Knauth, 1994; Prave, 1999; Corsetti and Kaufman, 2003; Macdonald and others, 2013b; Strauss and others, 2014a; Smith and others, 2015), coupled with these distinct formational boundaries, suggests a comparable sequence stratigraphic architecture for the Pahrump Gp. Notably, the sudden influx of fine-grained siliciclastic strata of unit KP1 into the Pahrump basin(s) is comparable in stratigraphic location (that is, post-dates the ICIE) to the T-R9 and T-R10 fault-related sedimentation in the Callison Lake and Kuujjua formations of NW Canada.

Mid-Neoproterozoic Tectonic Evolution of NW Canada and Laurentia

The Laurentian paleocontinent holds an analogous central position in Rodinia to Africa in Pangea (for example, Hoffman, 1991) – both share protracted extensional histories with the development of continent fringing passive margins and distinct relationships with plume-related continental flood basalt volcanism. Gondwana's break-up was characterized by the development of extensive, plume-related continental flood basalts (for example, Cox, 1978; Morgan, 1981; Encarnación and others, 1996) whose

1183 impact on the greater extensional history is still debated (Courtillot and others, 1999).
1184 Interestingly, this is very similar to the relationship between Neoproterozoic LIPs and the
1185 break up of Rodinia (for example, Li and others, 1999; 2008; Macdonald and others,
1186 2012; Sandeman and others, 2014; Yonkee and others, 2014).

1187 The widespread development of a prominent subaerial unconformity, volcanism,
1188 and rift-related sedimentation at the Mackenzie Mountains-Windermere Supergroup
1189 boundary in NW Canada appears to be contemporaneous with the emplacement of the
1190 Gunbarrel LIP around 780 Ma (Armstrong and others, 1982; Park and others, 1995;
1191 Dudás and Lustwerk, 1997; Harlan and others, 2003; Ootes and others, 2008; Sandeman
1192 and others, 2014). The Fifteenmile-Mount Harper Gp boundary in Yukon may be a local
1193 manifestation of this event; however, there is no evidence for Gunbarrel magmatism in
1194 Yukon and the new Re-Os and U-Pb ages from the Mount Harper Gp suggest that
1195 syntectonic units record pronounced extensional tectonism and sedimentation from
1196 ~750–720 Ma. This apparent ~30 Ma gap between Gunbarrel magmatism and regional
1197 extension in Yukon raises the interesting possibility of important ~750-730 Ma
1198 Laurentian tectono-sedimentary events (Evanchick and others, 1984; Parrish and
1199 Scammell, 1988; McDonough and Parrish, 1991; Crowley, 1997; Karlstrom and others,
1200 2000) that are completely unrelated to the Gunbarrel LIP and perhaps even more
1201 important in the long-term evolution of the western margin of Laurentia.

1202 In the Coal Creek inlier, Mustard (1991) and Mustard and Roots (1997) noted the
1203 development of diachronous, opposing basin geometries between the NNE-oriented Seela
1204 Pass and SSW-directed Eagle Creek half-grabens. Mustard (1991) predicted a discrete
1205 NW-SE-oriented extensional accommodation zone between these sub-basins that was

also responsible for localizing Mount Harper volcanism, similar to the spatial relationship between volcanism and accommodation zones in the East African Rift (Bosworth, 1985; 1987; Ebinger, 1989). The existence of this transfer zone is supported by the dense concentration of NW-oriented mafic dikes along the trace of this structure (fig. 4), as well as its documented role in dictating sedimentation patterns in the Coal Creek inlier and influencing the rotational motion of the Callison Lake depocenter during the Callison Lake-Seela Pass transition (fig. 14C).

The general pattern of propagating half-graben, coupled with the development of oblique intervening accommodation zones, is characteristic of the principal displacement zone of strike-slip extensional systems (Christie-Blick and Biddle, 1985 and references therein). Oblique extension in the Coal Creek inlier is possibly supported by the narrow rhomboidal map patterns in Windermere Supergroup deposits, distinct latitudinal and longitudinal basin asymmetries, apparently rapid subsidence of fault-related deposits, and the localized development of unconformities and abrupt lateral facies change (fig. 4; Mustard and Roots, 1997). More regionally, the lack of evidence for thermally driven post-rift subsidence in any of the Windermere sedimentary successions, even including the informal upper group, is also a characteristic feature of strike-slip sedimentation (Reading, 1980; Mann and others, 1983; Christie-Blick and Biddle, 1985 and references therein). Although Mustard (1991) opposed a model of oblique extension in the Coal Creek inlier, many of his arguments were based on non-unique sedimentological observations, such as the predominance of conglomeratic fill and the consistent orientation of paleocurrents transverse to the basin margin.

1228 In the Hart River inlier, Abbott (1997) documented evidence for syntectonic
1229 Mount Harper Gp sedimentation and coeval E-W-oriented mid-Neoproterozoic normal
1230 and reverse faults. In the en echelon fault belt of the Callison Lake and Rae Creek faults
1231 (fig. 3), Abbott (1997) noted a prominent difference in the erosional level of south-
1232 dipping Proterozoic strata beneath the Cambro-Ordovician Bouvette Fm. This
1233 observation, coupled with the consistent south-dipping orientation of steeply inclined
1234 structures and their apparent overlap by Rapitan-equivalent strata requires some degree of
1235 post-Callison Lake, pre-Rapitan reverse offset (fig. 3). Extremely localized zones of m-
1236 to km-scale folding of Fifteenmile and Pinguicula units are also present in the vicinity of
1237 relatively undeformed younger Neoproterozoic strata (for example, Penetration Lake, fig.
1238 3), although it is currently unclear if these structures are related to Mesozoic deformation.
1239 The restriction of Mount Harper Gp strata to south of the Callison Lake Thrust, along
1240 with the omission of units and opposing displacements across the Marc Creek Fault,
1241 suggests that these two structures are Neoproterozoic graben-bounding faults that were
1242 reactivated during the Mesozoic (fig. 3; Abbott, 1997). Importantly, all of these structural
1243 and stratigraphic configurations could be generated in a strike-slip setting where both
1244 normal and reverse fault separations are present in the same evolving fault system (for
1245 example, Nilsen and McLaughlin, 1985). In summary, map patterns, sedimentological
1246 observations, and structural arguments from both the Coal Creek and Hart River inliers
1247 support a reconstruction of initial dip-slip or slight oblique extension during deposition of
1248 the Callison Lake Fm followed by localized transtension and transpression during the
1249 development of ~740–660(?) Ma Seela Pass–Rapitan basins. Following previous
1250 suggestions for a Proterozoic origin of the Dawson Fault (Thrust) (Roots and Thompson,

1992; Abbott, 1996), we argue that this long-lived structure represents a fundamental trace of an ancient strike-slip fault zone in Yukon (fig. 14C).

Eisbacher (1978; 1981) documented evidence for localized pre- to syn-Windermere Supergroup contractional deformation in the Wernecke inlier (fig. 2), which was characterized by WSW-directed thrust faults involving the Wernecke Supergroup and Pinguicula Gp and capped by massive alluvial fan conglomerates of the Rapitan and Hay Creek groups. In the same region, Thorkelson (2000) mapped a number of related W- and SW-oriented thrust faults and overturned folds in the Pinguicula and Hematite Creek groups that were clearly truncated by overlying Windermere Supergroup conglomerates; Thorkelson (2000) termed this compressional event the “Corn Creek Orogeny”. These observations, in combination with the radiating orientation of Windermere Supergroup-related normal fault geometries and the regional angular unconformity between the Mackenzie Mountains and Windermere supergroups, led Eisbacher (1977; 1978; 1981) to interpret the Wernecke inlier as a discrete transpressional zone in a larger Windermere dextral strike-slip system.

Significant mid-Neoproterozoic angular unconformities and localized contractional structures are also described in the Coates Lake and Rapitan groups of the Mackenzie Mountains and regionally associated with the “Hayhook Orogeny” of Young and others (1979), or “Hayhook extensional event” of Jefferson and Parrish (1989) (Aitken and Cook, 1974; Helmstaedt and others, 1979; Jefferson, ms, 1983). The Hayhook event is a local manifestation of basal Windermere Supergroup extensional tectonism and is defined by discrete extensional faults and rift-related sedimentation in the Coates Lake and Rapitan groups (Young and others, 1979; Jefferson, ms, 1983;

Jefferson and Parrish, 1989). Although Eisbacher (1981) suggested that the compressional structures of the Wernecke inlier were synchronous with this basal Windermere rift-related sedimentation throughout NW Canada, Thorkelson (2000) mapped E-W-trending normal faults attributed to the Hayhook event that apparently crosscut the older thrust faults of the Corn Creek Orogeny, providing ambiguity to the proposed temporal link between Windermere Supergroup extension and compression. Alternatively, the normal faults of Thorkelson (2000) could represent tear faults in lateral thrust ramp complexes (*sensu* Thomas, 1990) and Eisbacher's (1981) original interpretation could still hold. Unfortunately, the direct connection between the enigmatic Corn Creek Orogeny and the Hayhook extensional event in the Wernecke inlier remains ambiguous; however, the presence of discrete regions characterized by transpressional and transtensional tectonism throughout NW Canada lends credence to the tectonic reconstructions proposed by Eisbacher (1977; 1981). In fact, many workers have proposed models of mid-Neoproterozoic strike-slip motion along the ancestral Richardson Fault Array and during deposition of the Coates Lake and Rapitan groups (Bell, 1982; Norris, 1982; Jefferson, ms, 1983; Jefferson and Ruelle, 1986; Jefferson and Parrish, 1989; Aitken and McMechan, 1992; Abbott, 1996).

When extrapolated south through the remainder of the North American Cordillera, NW Canada highlights a much more complex Neoproterozoic tectonic framework than simple dip-slip extension in Laurentian intracratonic basins (for example, Yonkee and others, 2014). A number of permissible scenarios could rectify these along-strike inconsistencies, including a different tectonic setting for NW Canadian versus ChUMP basins in the actively extending margin, diachronous extension and basin development,

1297 the localization of Gunbarrel plume-related thermal anomalies and related subsidence,
1298 and/or a distinct transition in the style and orientation of extensional tectonism along the
1299 length of ancestral North America. The consistency in stratigraphic architecture,
1300 sequence stratigraphy, $\delta^{13}\text{C}_{\text{carb}}$ chemostratigraphy, and radiometric age constraints
1301 appears to rule out diachroneity as an explanation for the earliest stages of Windermere
1302 Supergroup basin development; therefore, we hypothesize that this requires latitudinal
1303 (and possibly longitudinal) variability in extensional regimes, consistent with
1304 synchronous E-W or SE-NW dip-slip ChUMP extension coupled with NW-SE strike-slip
1305 extension in NW Canada (in present coordinates). Regional differences in the geometry
1306 of the basins may be due to the local tectonic fabric in the underlying basement (for
1307 example, Lund and others, 2010 and references therein); particularly, the ChUMP basins
1308 are present along reactivated Mesoproterozoic basins (for example, Macdonald and
1309 others, 2013b). Whether or not oblique extension also played a role in the subsequent
1310 development of ~720–660 Ma extensional basins throughout the western U.S. and
1311 Canada remains to be substantiated, although the distinct lack of tilted strata, modest
1312 extension (~25-40%), narrow basin geometries, lack of substantial volcanism, local
1313 detrital zircon sources, variable tectonic subsidence patterns, and lack of broad thermal
1314 subsidence permits a different interpretation from the current models of Neoproterozoic
1315 pure shear extension (Lund, 2008; Turner and Long, 2008; Lund and others, 2010;
1316 Yonkee and others, 2014). We suggest that the zigzag geometry of the Cordilleran rift
1317 system is an inherent Paleozoic feature (Hansen and others, 1993; Cecile and others,
1318 1997) and need not have defined the extensional framework of western Laurentia for over
1319 200 million years.

Implications for Neoproterozoic Chemostratigraphy and the Origin of the Islay Carbon

Isotope Excursion (ICIE)

The $\delta^{13}\text{C}_{\text{carb}}$ profile from the Callison Lake Fm is bracketed with U-Pb and Re-Os geochronology and has been used to correlate among regional and global pre-Sturtian basins (Macdonald and others, 2010; Strauss and others, 2014a). The chemostratigraphic data presented herein confirm previously published isotopic results and further demonstrate that the Callison Lake $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ data are reproducible within individual members and along depositional strike. An additional feature brought out with this study is a negative $\delta^{13}\text{C}_{\text{carb}}$ interval in ~750 Ma Talc mb strata (figs. 6, 7). Smith and others (2015) recently reported a similar negative $\delta^{13}\text{C}_{\text{carb}}$ interval from the correlative Horse Thief Springs Fm of the Pahrump Gp of the southwest United States, and an equivalent shift towards depleted $\delta^{13}\text{C}_{\text{carb}}$ values is recorded in the lower Russøya Mb of the Elbrogreen Fm of Svalbard (Halverson and others, 2004; 2005).

Despite the possible correlation of the Talc mb $\delta^{13}\text{C}_{\text{carb}}$ profile with other Neoproterozoic successions, Talc mb carbon isotope data remain somewhat ambiguous due their heterogeneity, as well as the depositional setting and complex paragenetic history of host carbonate strata. Sabkhas are characterized by complex early diagenetic environments involving the widespread precipitation, dissolution, and replacement of carbonate and sulfate minerals (Butler and others, 1982; Schreiber and El Tabkah, 2000; Warren, 2006 and references therein); these reactions commonly involve remineralization of organic matter through microbial sulfate reduction coupled with authigenic carbonate precipitation (Pierre and Rouchy, 1988; Anadón and others, 1992; Kendall, 2001; Machel,

2001). Syndepositional recrystallization to more stable carbonate minerals (for example, Reid and Macintyre, 1998) and penecontemporaneous dolomitization in sabkha environments contributes to the homogenization of authigenic and primary DIC isotopic compositions. Mixed carbonate-evaporite successions are also commonly affected by extensive thermochemical sulfate reduction during burial dolomitization (Machel, 2001 and references therein). Therefore, although we focused our sampling on the thickest exposures of pure microbial dolostone (figs. 6, 7), we cannot rule out based on the depositional setting that some, if not all, of the isotopic heterogeneity recorded in the Talc mb is a product of these early- and late-stage diagenetic processes. Importantly, many of the Talc mb stratigraphic sections remain predominantly enriched in $\delta^{13}\text{C}_{\text{carb}}$ (fig. 6), and these data appear to be more consistent with $\delta^{13}\text{C}_{\text{carb}}$ values from the bounding Heterolithic and Ramp members, as well as other coeval global successions (Halverson, 2006). It is worth noting that carbonate strata of the Horse Thief Springs Fm were also deposited in a sabkha-like depositional setting (Mahon and others, 2014), so the depleted $\delta^{13}\text{C}_{\text{carb}}$ values reported by Smith and others (2015) could be a result of similar processes involving the recrystallization of authigenic and primary carbonate phases.

The pervasive enrichment of $\delta^{18}\text{O}_{\text{carb}}$ throughout the Talc and Ramp members (fig. 13) lends further geochemical support to sedimentological evidence of evaporitic conditions throughout the Coal Creek and Hart River inliers. Although the chemostratigraphic utility of $\delta^{18}\text{O}_{\text{carb}}$ data is controversial, distinct $\delta^{18}\text{O}_{\text{carb}}$ enrichments are common in modern and ancient restricted settings (Friedman, 1980; McKenzie, 1981; Gat and Bowser, 1991; Kah and others, 1999; Frank and Lyons, 2000; Kah, 2000; Jaffrés and others, 2007; Wilson et al., 2010). Early dolomitization in Precambrian carbonate

successions aids in the reduction of pore space (Tucker, 1982) and thereby may restrict subsequent fluid interactions; however, it is also likely that $\delta^{18}\text{O}$ enrichments in ancient dolomites may simply be a function of the most recent pore-fluid composition and its formation temperature during precipitation, as well as some contribution from calcite-dolomite equilibrium isotope fractionation (for example, Land, 1983). Therefore, although it is difficult to distinguish between possible primary $\delta^{18}\text{O}_{\text{carb}}$ data and those that are imparted during subsequent diagenetic overprinting, the consistently enriched $\delta^{18}\text{O}_{\text{carb}}$ data of the Callison Lake Fm is at least consistent with primary modification of episodically restricted basin waters.

Predominantly enriched Ramp mb $\delta^{13}\text{C}_{\text{carb}}$ values from the Coal Creek inlier contrast with correlative strata in the Hart River region that display multiple, isotopically depleted intervals (figs. 6, 7). Some of these negative $\delta^{13}\text{C}_{\text{carb}}$ data appear to be associated with distinct paleokarst horizons (for example, Mark Creek, section J907; fig. 7); however, calling upon meteoric diagenesis as a driver for these depleted carbon isotope data (for example, Allan and Matthews, 1992; Knauth and Kennedy, 2009; Swart and Kennedy, 2012) is difficult to reconcile with the lack of evidence for a Neoproterozoic terrestrial biomass capable of generating abundant ^{13}C -depleted, soil-derived CO_2 (Jones and others, 2015). Other explanations include: 1) early diagenetic reactions involving evaporite replacement, dolomitization, or *in situ* microbial anaerobic respiration of organic carbon, 2) localized delivery and oxidation of detrital organic carbon, or 3) post-depositional alteration associated with burial dolomitization. High-energy intertidal to supratidal deposits of the Ramp mb are unusual lithofacies for pronounced authigenic carbonate production associated with anaerobic remineralization

of organic matter (for example, Irwin and others, 1977; Reimers and others, 1996; Schrag and others, 2013; Sun and Turchyn, 2014). Furthermore, there is no sedimentological or petrographic evidence for a sustained detrital organic carbon flux into the Hart River basin during Ramp mb sedimentation; this scenario would also require localized organic matter oxidation and carbonate precipitation (for example, Lloyd, 1964; Patterson and Walter, 1994) to outpace buffering through air-sea gas exchange. Although no distinct evaporite textures have been recognized in the Ramp mb, the m-scale breccias could represent isolated dissolution of precursor evaporite deposits (Pope and Grotzinger, 2003) and the supratidal depositional setting of these strata support episodic evaporitic conditions. Many of these depleted $\delta^{13}\text{C}_{\text{carb}}$ data are also associated with m-scale intervals of massively recrystallized dolostone, the origins of which are most likely related to fabric destructive burial-zone diagenesis. Therefore, we hypothesize that the minor divergence in compositional data between the Hart River and Coal Creek inliers most likely reflects local, platform-scale processes involving the early- and post-depositional interaction of evaporite recrystallization, organic matter oxidation, and dolomitization.

From predominantly enriched data in the Ramp mb, $\delta^{13}\text{C}_{\text{carb}}$ values of the Transitional mb drop to a nadir of -5.8‰ in a prominent negative excursion that has been previously correlated to the global ICIE due to its reproducibility in multiple sections and its covariation with $\delta^{13}\text{C}_{\text{org}}$ (fig. 6; Macdonald and others, 2010; Strauss and others, 2014a). This pre-Sturtian carbon isotope excursion was first recognized beneath the glaciogenic Port Askaig Fm in the Islay (Lossit) Limestone of Scotland (Brasier and Shields, 2000; McCay and others, 2006; Prave and others, 2009). It has also been documented in the Black River Fm of Tasmania (Calver, 1998), Mwashia Subgroup of

1412 the Roan Gp in Zambia (Bull and others, 2011), Beck Spring Dolomite of Death Valley
1413 (Horodyski and Knauth, 1994; Prave, 1999; Corsetti and Kaufman, 2003; Macdonald and
1414 others, 2013b; Smith and others, 2015), bed group 19 of NE Greenland (Fairchild and
1415 others, 2000), Russøya Mb of the Elbobreen Fm of Svalbard (Halverson and others,
1416 2004; Hoffman and others, 2012), Coppercap Fm of the Mackenzie Mountains
1417 (Halverson, 2006; Rooney and others, 2014), and the Kilian Fm of Victoria Island (Jones
1418 and others, 2010; Prince, 2014; Thomson and others, 2015b). Previous explanations for
1419 the ICIE were mechanistically linked to global glaciation due to the presence of
1420 isotopically depleted carbonate strata directly beneath circa 716-660 Ma Sturtian glacial
1421 deposits (Hoffman and others, 2012 and references therein); however, this relationship
1422 has recently been severed by syn-Islay Re-Os ages of 739.9 ± 6.1 Ma and 732.2 ± 3.9 Ma
1423 from the Callison Lake and Coppercap formations, respectively, which demand >10 Myr
1424 between the termination of the $\delta^{13}\text{C}_{\text{carb}}$ excursion and onset of glaciation (fig. 15; Rooney
1425 et al., 2014; Strauss et al., 2014a). These new ages are also consistent with the
1426 observation that many pre-glacial carbonate successions, including the Callison Lake Fm,
1427 contain a recovery to enriched $\delta^{13}\text{C}_{\text{carb}}$ values predating the onset of glaciation (fig. 6;
1428 Prave and others, 2009; Hoffman and others, 2012; Strauss and others, 2014a). Although
1429 a number of recent studies invoke diagenetic models for high amplitude Neoproterozoic
1430 $\delta^{13}\text{C}_{\text{carb}}$ excursions (for example, Knauth and Kennedy, 2009; Derry, 2010; Swart and
1431 Kennedy, 2012), Hoffman and others (2012) documented $\delta^{13}\text{C}_{\text{carb}}-\delta^{13}\text{C}_{\text{org}}$ covariation
1432 and a lack of $\delta^{18}\text{O}_{\text{carb}}$ covariance in the ICIE of Svalbard as evidence for a primary
1433 seawater DIC origin. These same isotopic covariance relationships are also present in the

Beck Spring Dolomite and Coppercap and Callison Lake formations (Corsetti and Kaufman, 2003; Rooney and others, 2014; Strauss and others, 2014a).

The relationship documented herein between circa 740 Ma tectonism, marine transgression, and the ICIE in Laurentia is consistent with a mechanistic link between tectonics, weathering, ocean geochemistry, and relative sea level change. Halverson and others (2014) highlighted the possible importance of Neoproterozoic continental flood basalt weathering in modulating the long-term carbon cycle during the protracted ca. 830–720 Ma low-latitude break-up of Rodinia. Here, we explore another feature of this pre-Sturtian tectonic background condition – the development and subsequent demise of extensive shallow epicontinental seaways and evaporite basins.

Despite the lack of consensus on the exact arrangement of paleocontinents in Rodinia (for example, Li and others, 2008; Evans, 2009), every hypothetical reconstruction necessitates significant marine extensional basins situated between larger continental fragments (Li and others, 2013) and paleocontinents that were episodically submerged beneath massive epicontinental seaways (for example, Centralian Superbasin of Australia; Walter and others, 1995; Lindsay, 2002). As noted above, the western margin of Laurentia was characterized by a series of extensional basins that were episodically linked to interior evaporitic epicontinental seaways such as the Amundsen Basin of Victoria Island (Young, 1981). Pre-Sturtian evaporite deposition in NW Canadian basins spans two discrete intervals (~900–811 Ma Minto Inlet-Ten Stone and ~780–740 Ma Kilian-Redstone River-Callison Lake) over an interpolated basin size $>3 \times 10^6 \text{ km}^2$ (Evans, 2006 and references therein). The coeval Centralian Superbasin of central Australia and Adelaide foldbelt of southern Australia represent an intermittently-

linked, extensive epicontinental seaway on the order of $2 \times 10^6 \text{ km}^2$ (Lindsay, 2002) that was also characterized by two protracted phases of coeval evaporite deposition ($>802 \text{ Ma}$ Curdimurka-Bitter Springs-Browne-Sunbeam and $\sim 780\text{--}720 \text{ Ma}$ Skillogalee) (Hill and Walter, 2000; Lindsay, 2002; Grey and others, 2011). Other volumetrically significant Neoproterozoic evaporitic basins include the $\sim 880(?)\text{--}720 \text{ Ma}$ Roan Gp of the central-African Copperbelt covering an estimated $0.5 \times 10^6 \text{ km}^2$ (Jackson and others, 2003; Armstrong and others, 2005; Selley and others, 2005; Bull and others, 2011) and the circa $800\text{--}720 \text{ Ma}$ Duruchaus Fm of the Kalahari craton that covered an estimated $0.3 \times 10^6 \text{ km}^2$ (Evans, 2006; Miller, 2008 and references therein). These enormous evaporite basins likely represent mere fractions of much larger epicontinental seaways that once existed between Rodinian paleocontinental fragments (for example, Li and others, 2013), many of which were eventually tectonically dismembered during the fragmentation of Rodinia.

How might the development and demise of these epicontinental seaways drive the ICIE and affect the long-term Neoproterozoic carbon cycle? If the ICIE is analogous to the Paleocene-Eocene Thermal Maximum (PETM) negative $\delta^{13}\text{C}_{\text{carb}}$ excursion and lasted for a duration comparable to the mixing and residence times of carbon in the oceans (1–10s kyr), then a plausible driving mechanism involves the extensive subaerial oxidation of organic matter associated with the uplift and erosion of these evaporite basins (*cf.* Higgins and Schrag, 2006). If, however, the ICIE turns out to have lasted on a timescale greater than the residence time of carbon in the oceans ($>100 \text{ kyr}$), a massive carbon oxidation explanation is less likely given the required oxidant budgets (for example, Bristow and Kennedy, 2008) and scenarios involving changes in the fractional burial of

organic carbon (Kump, 1991) and authigenic carbonate (Schrag and others, 2013) or shifts in the composition of weathering products (Kump and others, 1999) provide more reasonable explanations. Epicontinental seaways characterized by extensive sabkha and basinal evaporite deposits not only represent an important sink for authigenic carbonate and organic carbon burial, but also act as critical regulators of marine sulfate inventories during precipitation and dissolution events (for example, Wortmann and Paytan, 2012).

Similar to records of enhanced organic matter preservation in large Mesozoic epicontinental seaways (for example, Slingerland and others, 1996; Fisher and Arthur, 2002; Riboulleau and others, 2003), Rodinian extensional basins likely provided a locus for elevated organic matter and authigenic carbonate burial, as evidenced by the consistently enriched background $\delta^{13}\text{C}_{\text{carb}}$ data in the early Neoproterozoic (for example, Knoll and others, 1986; Schrag and others, 2002; Halverson and others, 2005). As an example, evaporitic strata in the Talc mb of the Callison Lake Fm contain organic-rich black shale horizons with up to 4 wt.% total organic carbon in association with authigenic carbonate with depleted $\delta^{13}\text{C}_{\text{carb}}$ values (figs. 6, 7). On short timescales, the desiccation of even some small fraction of these massive epicontinental seaways during the discrete ~740 Ma episode of regional extension discussed above would have resulted in the widespread oxidation of labile marine organic carbon through interactions with oxygenated meteoric systems, sulfate-rich brines, and exposure to aerobic microbial respiration (for example, Hartnett and others, 1998; Hedges and others, 1999; Moodley and others, 2005; Bouchez and others, 2010). Higgins and Schrag (2006) estimated that assuming >90% of the total sedimentary organic carbon is oxidized during an episode of uplift and exposure, the desiccation of the top ~30 m of an $\sim 3 \times 10^6 \text{ km}^2$ epicontinental

seaway with 5 wt.% organic carbon would result in the release of ~5000 Gt C to the ocean-atmosphere system. This hypothetical seaway is essentially the same size as only one of the major *preserved* Rodinian epicontinental basins, the majority of which display abundant sedimentological evidence for sub-Sturtian erosional unconformities (for example, Jefferson, ms, 1983; Rainbird, 1993; Prave and others, 2009; Hoffman and others, 2012 and many others).

Given this framework and based on simple mass balance calculations considering the modern ocean-atmosphere system, the approximate amount of oxidized organic carbon ($\delta^{13}\text{C} = -25\text{‰}$) required to generate a ~10‰ negative $\delta^{13}\text{C}_{\text{carb}}$ excursion is ~22,000 Gt. This calculation neglects effects such as CaCO_3 dissolution and carbonate speciation that would likely require even larger initial carbon inputs; however, the exact magnitude of the required Islay-related CO_2 flux also depends on many critical features that remain poorly constrained in the Neoproterozoic, such as the size of the ocean-atmosphere-biosphere carbon pool, the composition of seawater (Ca^{2+} , alkalinity, et cetera), the attendant distribution of CO_2 throughout the carbonate system, and its final consumption by biological productivity. One would assume that the uplift of Rodinian evaporite basins and rift flanks would have also initiated weathering of sulfate evaporites and remnants of Neoproterozoic continental flood basalt provinces, thereby possibly limiting the total CO_2 release generated by widespread organic matter oxidation through enhanced nutrient delivery and biological uptake of CO_2 (for example, Tziperman and others, 2011).

One prediction of this organic matter oxidation model is that the ICIE should be marked by rapid warming and sea level rise due to the thermal expansion of seawater. The consistent association of the ICIE with marine transgression in extensional basins of

Laurentia (fig. 12) provides preliminary support for a hypothetical flooding event, which may be mirrored in other global successions (for example, Prave and others, 2009; Bull and others, 2011; Hoffman and others, 2012; Smith and others, 2015); however, the Laurentian sea level record is difficult to disentangle from extensional tectonism as a primary driver for regional base level rise (see above). Given the predicted sea level response, this hypothesis also requires rapid uplift rates to expose evaporite basins while simultaneously isolating them from warming-related marine transgression (Higgins and Schrag, 2006). Although this may appear challenging to explain in extensional settings, the combined effects of plume-related uplift (Saunders and others, 2007), dynamic topography, and oblique extension during the break-up of Rodinia may have promoted the widespread development of basinal sills (for example, Duggen and others, 2003) and facilitated relatively rapid (for example, Vogl and others, 2014) isolation, uplift, and weathering of pre-Sturtian evaporite basins.

The observation in the Callison Lake Fm and other global successions that the ICIE spans 10s of m of mixed carbonate and siliciclastic strata and contains variable internal structure (figs. 6, 15) possibly supports a longer duration for the $\delta^{13}\text{C}_{\text{carb}}$ excursion. In order to circumvent oxidant mass balance complications associated with driving long lived $\delta^{13}\text{C}_{\text{carb}}$ excursions with organic matter burial, Schrag and others (2013) proposed that authigenic carbonate production provides a sink for ^{13}C -depleted carbon in Proterozoic oceans characterized by low O_2 and alkalinity-primed anoxic pore fluids (for example, Higgins and others, 2009). Due to the relatively consistent magnitude of the excursion and its $\delta^{13}\text{C}_{\text{carb}}-\delta^{13}\text{C}_{\text{org}}$ covariation, the ICIE could be related to a global decline in the fractional burial of authigenic carbonate (category 1 of Schrag and others,

2013) due to a transient perturbation to surface redox conditions or seawater carbonate saturation state, an abrupt sea level fluctuation, and/or a shift in the loci of global organic carbon remineralization in marine basins. The margin-wide ~ 740 Ma extensional episode and marine transgression in Laurentia may have temporarily altered the locus of sediment column organic carbon remineralization (for example, by shoaling the zone of anaerobic oxidation of methane) or shifted the primary location of organic carbon burial and remineralization in Rodinian extensional basins. One criticism of the authigenic carbonate model is the difficulty in identifying the exact nature and location of the depleted carbonate sink – significant volumes of sulfate and authigenic carbonate could have been sequestered in the vast epicontinental basins of Rodinia. Therefore, the episodic dissection of these basins during regional tectonic events may have delivered light carbon and sulfate to the oceans through a combination of direct oxidation of organic matter and weathering of authigenic carbonate reservoirs, while simultaneously affecting the sinks of authigenic carbonate and organic carbon burial in isolated extensional basins.

Sulfur isotope data support a hypothetical evaporite weathering hypothesis for the ICIE. Sulfur isotopic data from pre-Sturtian sedimentary deposits display profound variability (for example, Gorjan and others, 2000; Hurtgen and others, 2002; Halverson and Hurtgen, 2007), consistent with low marine SO_4^{2-} concentrations possibly associated with ca. 850-720 Ma global evaporative drawdown. Moreover, $\delta^{34}\text{S}$ data from sulfate deposits of the Kilian Fm record a ~15‰ positive anomaly that covaries with the ICIE (Kaufman and others, 2007; Jones and others, 2010), which is consistent with the dissolution of freshly exposed sulfate evaporites and attendant increase in pyrite burial

through elevated microbial sulfate reduction (*sensu* Wortmann and Paytan, 2012). Set upon a backdrop of low latitude continental configurations, the coincidence in timing of other high amplitude Neoproterozoic $\delta^{13}\text{C}_{\text{carb}}$ anomalies (for example, ca. 811 Ma Bitter Springs anomaly) with distinct episodes of intra-Rodinian extensional tectonism suggests that this evaporative drawdown and weathering hypothesis may have far reaching implications for Neoproterozoic global climate and biogeochemical cycles.

CONCLUSIONS

The Callison Lake Fm (formalized herein) of the Coal Creek and Hart River inliers of Yukon, Canada, records a complex subsidence history, in which episodic basinal restriction and abrupt facies change can be tied to mid-Neoproterozoic (Tonian) extensional tectonism and rift-related sedimentation (Windermere Supergroup) throughout NW Canada. The syn-tectonic, circa 753 Ma Heterolithic mb of the basal Callison Lake Fm is dominated by fine-grained siliciclastic sedimentation in a marginal marine depositional setting and locally includes discontinuous stromatolitic bioherms with poorly preserved VSMs. These strata are sharply overlain by a progradational sabkha succession (Talc mb) dominated by talc-rich black shale interbedded with evaporitic microbial dolostone, which transition upsection into a thick package of subtidal to supratidal dolostone (Ramp mb) and mark the development of a carbonate ramp. In some stratigraphic sections in the Coal Creek inlier, Ramp mb carbonate strata grade up-section into interbedded microbial dolostone and organic-rich black shale deposits of the Transitional mb, which contains a diverse VSM assemblage constrained by a Re-Os depositional age of 739.9 ± 6.1 Ma. In other stratigraphic sections, this same

depositional interval is marked by a significant paleokarst horizon, which is suggestive of differential subsidence and uplift associated with a major phase of ~740 Ma extensional tectonism in Yukon.

Stratigraphic, structural, and geochronological data from the Callison Lake Fm demonstrate that these strata are correlative with at least part of the ~780–720 Ma Coates Lake Gp of the Mackenzie Mountains, both of which were potentially deposited during ~740–720 Ma transtension and transpression along the northwestern margin of Laurentia. New sequence stratigraphic data described herein from the Callison Lake Fm highlights three distinct depositional sequences, or transgressive-regressive (T-R) cycles, that are coeval with similar stratigraphic packages in the Coates Lake Group of the Mackenzie Mountains, Shaler Supergroup of Victoria Island, and Chuar-Uinta Mountain-Pahrump basins of the western United States. The global, circa 735 Ma Islay carbon isotope excursion (ICIE) is consistently present in the third T-R cycle and is interpreted to represent a primary perturbation to the global carbon cycle based on its reproducibility in regional and global basins and covariance in carbonate and organic carbon isotopes. Here, we explore a new model for the origin of the ICIE that links this negative carbon isotope excursion to the uplift and weathering of extensive shallow epicontinental seaways and evaporite basins associated with the break-up of Rodinia.

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REFERENCES

- Abbott, J. G., 1993, Revised stratigraphy and new exploration targets in the Hart River area (NTS 116A/10, 116A/11), southeastern Yukon, *in* Yukon Exploration and Geology 1992: Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, p. 13-23.
- Abbott, J. G., 1996, Implications of probable Late Proterozoic dextral strike-slip movement on the Snake River Fault, *in* Cook, F.A. and Erdmer, P., editors, Slave – Northern Cordillera Lithospheric Evolution (SNORCLE) Transect: Lithoprobe Report 50, p. 138-140.
- Abbott, G., 1997, Geology of the upper Hart River area, eastern Ogilvie Mountains, Yukon Territory (116A/10, 116A/11): Exploration and Geological Services Division, Yukon Region, Bulletin 9, p. 1-76.
- Abercombie, H.J., ms, 1978, Mineralogy and environment of deposition of a Precambrian evaporite deposit, Mackenzie Mountains, Northwest Territories: B.Sc. thesis, University of British Columbia, 45 p.

1650
1651 Aitken, J.D., 1981, Stratigraphy and sedimentology of the upper Proterozoic Little Dal Group,
1652 Mackenzie Mountains, Northwest Territories, *in* Campbell, F.H.A., editor, Proterozoic
1653 basins of Canada: Geological Survey of Canada Paper 81-10, p. 47-71.
1654
1655 Aitken, J.D., 1991, The Ice Brook Formation and post-Rapitan, Late Proterozoic glaciation,
1656 Mackenzie Mountains, Northwest Territories: Geological Survey of Canada Bulletin 404,
1657 43 p.
1658
1659 Aitken, J.D., and Cook, D.G., 1974, Geological maps, northern part of Mount Eduni and Bonnet
1660 Plume Lake map-areas, District of Mackenzie: Geological Survey of Canada Open File
1661 Report 221, scale 1:250,000.
1662
1663 Aitken, J.D., and McMechan, M.E., 1992, Middle Proterozoic Assemblages, *in* Gabrielse, H., and
1664 Yorath, C.J., editors, Geology of the Cordilleran Orogen in Canada: Geological Survey of
1665 Canada, Geology of Canada, n. 4, p. 97-124. (also Geological Society of America, The
1666 Geology of North America, v. G-2).
1667
1668 Allen, P.A., and Homewood, P., 1984, Evolution and mechanics of a Miocene tidal dune:
1669 Sedimentology, v. 31, p. 62-81.

1670 Anadón, P., Rosell, L., and Talbot, M.R., 1992, Carbonate replacement of lacustrine gypsum
1671 deposits in two Neogene continental basins, eastern Spain: Sedimentary Geology, v. 78, p.
1672 201-216.

1673 Armin, R.A., and Mayer, L., 1983, Subsidence analysis of the Cordilleran miogeocline:
1674 Implications for timing of late Proterozoic rifting and amount of extension: Geology, v. 11,
1675 p. 702-705.

1676 Armstrong, R.A., Eisbacher, G.H., and Evans, P.D., 1982, Age and stratigraphic-tectonic
1677 significance of Proterozoic diabase sheets, Mackenzie Mountains, northwestern Canada:
1678 Canadian Journal of Earth Sciences, v. 19, no. 2, p. 316-323.

1679 Armstrong, R.A., Master, S., and Robb, L.J., 2005, Geochronology of the Nchanga Granite, and
1680 constraints on the maximum age of the Katanga Supergroup, Zambian Copperbelt: Journal
1681 of African Earth Sciences, v. 42, p. 32-40.

1682 Ashley, G.M., 1990, Classification of large scale subaqueous bedforms: a new look at an old
1683 problem: Journal of Sedimentary Petrology, v. 60, p. 160-172.
1684
1685 Bell, R.T., 1982, Comments on the geology and uraniferous mineral occurrences of the Wernecke
1686 Mountains, Yukon and District of Mackenzie: Current Research Part B, Geological Survey
1687 of Canada, Paper 1982-1B, p. 279-284.
1688
1689 Blaise, B., and Mercier, E., 1984, La sédimentation au Précambrien supérieur et au Paléozoïque
1690 inférieur sur la marge de la plate-forme du Mackenzie (Monts Ogilvie, Territoire du
1691 Yukon): Current Research Part B, Geological Survey of Canada, Paper 84-1B, p. 85-92.
1692
1693 Bond, G.C., and Kominz, M.A., 1984, Construction of tectonic subsidence curves for the early
1694 Paleozoic miogeocline, southern Canadian Rocky Mountains: implications for subsidence
1695 mechanisms, age of breakup, and crustal thinning: Geological Society of America Bulletin,

- 1696 v. 95, no. 2, p. 155.
- 1697 Bond, G.C., Kominz, M.A., and Devlin, W.J., 1983, Thermal subsidence and eustasy in the
1698 Lower Paleozoic miogeocline of western North America: *Nature*, v. 306, no. 5945, p. 775–
1699 779, doi: 10.1038/306775a0.
- 1700 Bond, G.C., Christie-Blick, N., Kominz, M.A., and Devlin, W.J., 1985, An early Cambrian rift to
1701 post-rift transition in the Cordillera of western North America: *Nature*, v. 315, no. 6022, p.
1702 742–746, doi: 10.1038/315742a0.
- 1703 Bosence, D.W.J., 1998, Stratigraphic and sedimentological models of rift basins, *in* Purser, B.H.,
1704 and Bosence, D.W.J., editors, *Sedimentation and Tectonics of Rift Basins: Red Sea–Gulf of*
1705 *Aden*: London, Chapman and Hall, p. 9-28.
- 1706
- 1707 Bosworth, W., 1985, Geometry of propagating continental rifts: *Nature*, v. 316, p. 625-627.
- 1708
- 1709 Bosworth, W., 1987, Off-axis volcanism in the Gregory rift, east Africa: Implications for models
1710 of continental rifting: *Geology*, v. 15, p. 397-400.
- 1711 Bosworth, W., Crevello, P., Winn Jr., R.D., and Steinmetz, J., 1998, Structure, sedimentation, and
1712 basin dynamics during rifting of the Gulf of Suez and northwestern Red Sea, *in* Purser,
1713 B.H., and Bosence, D.W.J., editors, *Sedimentation and Tectonics of Rift Basins: Red Sea–*
1714 *Gulf of Aden*: London, Chapman and Hall, p. 77-96.
- 1715
- 1716 Bosworth, W., Huchon, P., and McClay, K., 2005, The Red Sea and Gulf of Aden Basins: *Journal*
1717 *of African Earth Sciences*, v. 43, no. 1–3, p. 334–378.
- 1718 Bowring, S.A., and Grotzinger, J.P., 1992, Implications of new chronostratigraphy for tectonic
1719 evolution of Wopmay Orogen, northwest Canadian Shield: *American Journal of Science*, v.
1720 292, p. 1-20.
- 1721
- 1722 Braiser, M.D., and Shields, G., 2000, Neoproterozoic chemostratigraphy and correlation of the
1723 Port Askaig glaciation: Dalradian Supergroup of Scotland: *Journal of the Geological*
1724 *Society of London*, v. 157, p. 909-914.
- 1725 Bristow, T.F., and Kennedy, M.J., 2008, Carbon isotope excursions and the oxidant budget of the
1726 Ediacaran atmosphere and ocean: *Geology*, v. 36, n. 11, p. 863-866.
- 1727 Bull, S., Selley, D., Broughton, D., Hitzman, M., Cailteux, J., Large, R., and McGoldrick, P.,
1728 2011, Sequence and carbon isotopic stratigraphy of the Neoproterozoic Roan Group strata
1729 of the Zambian copperbelt: *Precambrian Research*, v. 190, p. 70-89.
- 1730
- 1731 Butler, G.P., Harris, P.M., and Kendall, C.G. St.C., 1982, Recent evaporites from the Abu Dhabi
1732 coastal flats, *in* Hanford, C.R., Loucks, R.G., and Davies, G.R., editors, *Deposition and*
1733 *Diagenetic Spectra of Evaporites*: Society of Economic Paleontologists and Mineralogists
1734 (SEPM) Core Workshop, v. 3, p. 33–64.
- 1735
- 1736 Calver, C.R., 1998, Isotope stratigraphy of the Neoproterozoic Togari Group, Tasmania:
1737 *Australian Journal of Earth Sciences*, v. 45, p. 865-874.
- 1738 Calvo, J.P., Blanc-Valleron, M.M., Rodriguez-Arandia, J.P., Rouchy, J.M., and Sanz, M.E., 1999,

- 1739 Authigenic clay minerals in continental evaporitic environments: Special Publication of the
1740 International Associations of Sedimentologists, v. 27, p. 129–152.
- 1741 Carr, I.D., Gawthorpe, R.L., Jackson, C.A.L., Sharp, I.R., and Sadek, A., 2003, Sedimentology
1742 and sequence stratigraphy of early syn-rift tidal sediments: The Nukhul Formation, Suez
1743 Rift, Egypt: *Journal of Sedimentary Research*, v. 73, n. 3, p. 407–420.
- 1744 Cattaneo, A., and Steel, R.J., 2003, Transgressive deposits: a review of their variability: *Earth-*
1745 *Science Reviews*, v. 62, p. 187–228.
- 1746 Catuneanu, O., Abreu, V., Bhattacharya, J.P., Blum, M.D., Dalrymple, R.W., Eriksson, P.G.,
1747 Fielding, C.R., Fisher, W.L., Galloway, W.E., Gibling, M.R., Giles, K.A., Holbrook, J.M.,
1748 Jordan, R., Kendall, C.G.St.C., Macurda, B., Martinsen, O.J., Miall, A.D., Neal, J.E.,
1749 Nummedal, D., Pomar, L., Posamentier, H.W., Pratt, B.R., Sarg, J.F., Shanley, K.W., Steel,
1750 R.J., Strasser, A., Tucker, M.E., and Winker, C., 2009, Towards the standardization of
1751 sequence stratigraphy: *Earth-Science Reviews*, v. 92, p. 1–33.
- 1752 Catuneanu, O., Galloway, W.E., Kendall, C.G.St.C., Miall, A.D., Posamentier, H.W., Strasser, A.,
1753 and Tucker, M.E., 2011, Sequence Stratigraphy: Methodology and Nomenclature:
1754 *Newsletters on Stratigraphy*, v. 44, no. 3, p. 173–245, doi: 10.1127/0078-0421/2011/0011.
- 1755 Cecile, M.P., Morrow, D.W., and Williams, G.K., 1997, Early Paleozoic (Cambrian to Early
1756 Devonian) tectonic framework, Canadian Cordillera: *Bulletin of Canadian Petroleum*
1757 *Geology*, v. 45, p. 54–74.
- 1758
1759 Cerling, T.E., 1994, Chemistry of closed basin lake waters: A comparison between African Rift
1760 Valley and some North American rivers and lakes, *in* Gierlowski-Kordesch, E., and Kelts,
1761 K., eds., *Global Geological Record of Lake Basins*: IGCP Monograph 219, v. 1, p. 29–30.
1762
- 1763 Christie-Blick, N., and Biddle, K.T., 1985, Deformation and basin formation along strike-slip
1764 faults, *in* Biddle, K.T., and Christie-Blick, N., editors, *Strike-Slip Deformation, Basin*
1765 *Formation, and Sedimentation*: Society of Economic Paleontologists and Mineralogists
1766 (SEPM) Special Publication, v. 37, p. 1–34.
1767
- 1768 Cohen, P.A., and Macdonald, F.A., 2015, The Proterozoic record of eukaryotes: *Paleobiology*, v.
1769 XX, p. 1–23.
1770
- 1771 Colpron, M., Logan, J., and Mortensen, J., 2002, U-Pb zircon age constraint for late
1772 Neoproterozoic rifting and initiation of the lower Paleozoic passive margin of western
1773 Laurentia: *Canadian Journal of Earth Sciences*, v. 39, no. 2, p. 133–143, doi: 10.1139/E01-
1774 069.
- 1775 Cook, F.A., and MacLean, B.C., 1995, The intracratonic Paleoproterozoic Forward orogeny, and
1776 implications for regional correlations, Northwest Territories, Canada: *Canadian Journal of*
1777 *Earth Science*, v. 32, p. 1991–2008.
- 1778 Cook, F.A., Varsek, J.L., and Clark, E.A., 1991, Proterozoic craton to basin crustal transition in
1779 western Canada and its influence on the evolution of the Cordillera: *Canadian Journal of*
1780 *Earth Sciences*, v. 28, no. 8, p. 1148–1158.
- 1781 Cook, F.A., Dredge, M., and Clark, E.A., 1992, The Proterozoic Fort Simpson structural trend in

- 1782 northwestern Canada: Geological Society of America Bulletin, v. 104, p. 1121-1137.
- 1783 Corsetti, F.A., and Kaufman, A.J., 2003, Stratigraphic investigations of carbon isotope anomalies
1784 and Neoproterozoic ice ages in Death Valley, California: Geological Society of America
1785 Bulletin, v. 115, n. 8., p. 916-932.
- 1786 Courtillot, V., Jaupart, C., Manighetti, P., Tapponnier, P., and Besse, J., 1999, On causal links
1787 between flood basalts and continental breakup: Earth and Planetary Science Letters, v. 166,
1788 p. 177-195.
- 1789 Cox, G.M., Roots, C.F., Halverson, G.P., Minarik, W.G., Macdonald, F.A., and Hubert-Theou, L.,
1790 2013, Mount Harper Volcanic Complex, Ogilvie Mountains: A far-flung occurrence of the
1791 Franklin Igneous Event?, *in* MacFarlane, K.E., Nordling, M.G., and Sack, P.J., editors,
1792 Yukon Exploration and Geology 2012: Yukon Geological Survey, p. 19-36.
- 1793 Cox, K.G., 1978, Flood basalts, subduction and the break-up of Gondwanaland: Nature, v. 274, p.
1794 47-49.
- 1795 Crittenden, M.D., Jr., Schaeffer, F.E., Trimble, D.E., and Woodward, L.A., 1971, Nomenclature
1796 and correlation of some upper Precambrian and basal Cambrian sequences in western Utah
1797 and southeastern Idaho: Geological Society of America Bulletin, v. 94, p. 437-450.
- 1798
1799 Crowley, J.L., 1997, U-Pb geochronologic constraints on the cover sequence of the Monashee
1800 complex, Canadian Cordillera: Paleoproterozoic deposition on basement: Canadian Journal
1801 of Earth Science, v. 34, p. 1008-1022.
- 1802
1803 Dalrymple, R.W., 2010, Tidal depositional systems, *in* James, N.P., and Dalrymple, R.W., editors,
1804 Facies Models 4: Geological Association of Canada GEOText 6, p. 201-231.
- 1805
1806 Dehler, C.M., Fanning, C.M., Link, P.K., Kingsbury, E.M., and Rybczynski, D., 2010, Maximum
1807 depositional age and provenance of the Uinta Mountain Group and Big Cottonwood
1808 Formation, northern Utah: Paleogeography of rifting western Laurentia: Geological Society
1809 of America Bulletin, v. 122, no. 9-10, p. 1686–1699, doi: 10.1130/B30094.1.
- 1810 Dehler, C.M., Karlstrom, K.E., Gehrels, G.E., Timmons, M.J., and Crossey, L.J., 2012,
1811 Stratigraphic revision, provenance, and new age constraints of the Nankoweap Formation
1812 and Chuar Group, Grand Canyon Supergroup, Grand Canyon, Arizona: Geological Society
1813 of America Abstracts with Programs, v. 44, no. 6, p. 82.
- 1814
1815 Dekov, V.M., Cuadros, J., Shanks, W.C., and Koski, R.A., 2008, Deposition of talc, kerolite-
1816 smectite, smectite at seafloor hydrothermal vent fields: Evidence from mineralogical,
1817 geochemical and oxygen isotope studies: Chemical Geology, v. 247, p. 171–194.
- 1818 Delaney, G.D., 1981, The mid-Proterozoic Wernecke Supergroup, Wernecke Mountains, Yukon
1819 Territory, *in* Campbell, F.H.A., editor, Proterozoic Basins of Canada: Geological Survey of
1820 Canada Paper 81-10, p. 1-23.
- 1821
1822 Desjardins, P.R., Buatois, L.A., and Mangano, M.G., 2012, Tidal flats and subtidal sand bodies,
1823 *in* Knaust, D., and Bromley, R.G., editors, Trace fossils as indicators of sedimentary
1824 environments: Developments of Sedimentology, v. 64, p. 529-561.

- 1825 Devlin, W.J., and Bond, G.C., 1988, The initiation of the early Paleozoic Cordilleran
1826 miogeocline: evidence from the uppermost Proterozoic-Lower Cambrian Hamill Group of
1827 southeastern British Columbia: Canadian Journal of Earth Sciences, v. 25, no. 1, p. 1-19.
- 1828 Donnadieu, Y., Godderis, Y., Ramstein, G., Nedelec, A., and Meert, J., 2004, A 'snowball Earth'
1829 climate triggered by continental break-up through changes in runoff: Nature, v. 428, p.
1830 303-306.
- 1831 Dudás, F.Ö., and Lustwerk, R.L., 1997, Geochemistry of the Little Dal basalts: continental
1832 tholeiites from the Mackenzie Mountains, Northwest Territories, Canada: Canadian Journal
1833 of Earth Science, v. 34, p. 50-58.
- 1834 Duggen, S., Hoernle, K., van den Bogaard, P., Rupke, L., and Morgan, J.P., 2003, Deep roots of
1835 the Messinian salinity crisis: Nature, v. 422, p. 602-605.
- 1836 Ebinger, C.J., 1989, Tectonic development of the western branch of the East African rift system:
1837 Geological Society of America Bulletin, v. 101, p. 885-903.
- 1838 Eisbacher, G.H., 1977, Tectono-stratigraphic framework of the Redstone Copper Belt, District of
1839 Mackenzie, Report of Activities Part A: Geological Survey of Canada, Paper 77-1A, p.
1840 229-234.
- 1841 Eisbacher, G.H., 1978, Two major Proterozoic unconformities, northern Cordillera, Current
1842 Research Part A: Geological Survey of Canada, Paper 78-1A, p. 53-58.
- 1843 Eisbacher, G.H., 1981, Sedimentary tectonics and glacial record in the Windermere Supergroup,
1844 Mackenzie Mountains, northwestern Canada: Geological Survey of Canada Paper 80-27,
1845 40 p.
- 1846 Eisbacher, G.H., 1985, Late Proterozoic rifting, glacial sedimentation, and sedimentary cycles in
1847 the light of Windermere deposition, western Canada: Palaeogeography, Palaeoclimatology,
1848 and Palaeoecology, v. 51, p. 231-254.
- 1849 Embry, A.F., 1995, Sequence boundaries and sequence hierarchies: problems and proposals, *in*
1850 Steel, R.J., Felt, V.L., Johannessen, E.P., and Mathieu, C., editors, Sequence stratigraphy
1851 on the Northwest European Margin: Norwegian Petroleum Society (NPF), Special
1852 Publication 5, p. 1-11.
- 1853 Embry, A.F., 2009, Practical Sequence Stratigraphy: Canadian Society of Petroleum Geologists,
1854 79 p., online at www.cspg.org
- 1855 Embry, A.F., and Johannessen, E.P., 1992, T-R sequence stratigraphy, facies analysis and
1856 reservoir distribution in the uppermost Triassic-Jurassic succession, western Sverdrup
1857 Basin, Arctic Canada, *in* Vorren, T.O., Bergsager, E., Dahl-Stamnes, O.A., Holter, E.,
1858 Johansen, B., Lie, E., and Lund, T.B., editors, Arctic Geology and Petroleum Potential,
1859 Norwegian Petroleum Society, Special Publication, v. 2, p. 121-146.
- 1860 Encarnación, J., Fleming, T.H., Elliot, D.H., Eales, H.V., 1996, Synchronous emplacement of the
1861 Ferrar and Karoo dolerites and the early breakup of Gondwana: Geology, v. 24, p. 535-538.

- 1862 Evanchick, C.A., Parrish, R.R., and Gabrielse, H., 1984, Precambrian gneiss and late Proterozoic
1863 sedimentation in north-central British Columbia: *Geology*, v. 12, p. 233-237.
1864
- 1865 Evans, D.A.D., 2006, Proterozoic low orbital obliquity and axial-dipolar geomagnetic field from
1866 evaporite palaeolatitudes: *Nature*, v. 444, p. 51-56.
1867
- 1868 Evans, D.A.D., 2009, The palaeomagnetically viable, long-lived and all-inclusive Rodinia
1869 supercontinent reconstruction: *Journal of the Geological Society of London, Special*
1870 *Publications*, v. 327, p. 371-404.
- 1871 Fairchild, I.J., Spiro, B., Herrington, P.M., and Song, T.R., 2000, Controls on Sr and C isotopic
1872 compositions of Neoproterozoic Sr-rich limestones of East Greenland and North China, *in*
1873 Grotzinger, J.P., and James, N.P., editors, *Carbonate Sedimentation and Diagenesis in the*
1874 *Evolving Precambrian World*, Society of Economic Paleontologists and Mineralogists
1875 (SEPM) Special Publication, v. 67, p. 297-313.
- 1876 Fisher, C.G., and Arthur, M.A., 2002, Water mass characteristics in the Cenomanian U.S.
1877 Western Interior Seaway as indicated by stable isotopes of calcareous organisms:
1878 *Palaeogeography, Palaeoclimatology, and Palaeoecology*, v. 188, p. 189-213.
1879
- 1880 Folk, R.L., and Pittman, J.S., 1971, Length-slow chalcedony: A new testament for vanished
1881 evaporites: *Journal of Sedimentary Petrology*, v. 41, no. 4, p. 1-14.
- 1882 Frank, T.D., and Lyons, T.W., 2000, The integrity of $\delta^{18}\text{O}$ records in Precambrian carbonates: A
1883 Mesoproterozoic case study, *in* Grotzinger, J.P., and James, N.P., editors, *Carbonate*
1884 *Sedimentation and Diagenesis in the Evolving Precambrian World: Society for*
1885 *Sedimentary Geology (SEPM) Special Publication*, v. 67, p. 315-326.
- 1886 Friedman, G.M., 1980, Dolomite is an evaporite mineral: evidence from the rock record and from
1887 sea-marginal ponds of the Red Sea, *in* Zenger, D.H., Dunham, J.D., and Ethington, R.L.,
1888 editors, *Concepts and Models of Dolomitization: Society of Economic Paleontologists and*
1889 *Mineralogists (SEPM) Special Publication*, v. 28, p. 69-80.
1890
- 1891 Furlanetto, F., Thorkelson, D.J., Gibson, H.D., Marshall, D.D., Rainbird, R.H., Davis, W.J.,
1892 Crowley, J.L., and Vervoort, J.D., 2013, Late Paleoproterozoic terrane accretion in
1893 northwestern Canada and the case for circum-Columbian orogenesis: *Precambrian*
1894 *Research*, v. 224, p. 512-528.
1895
- 1896 Gabrielse, H., 1967, Tectonic evolution of the northern Canadian Cordillera: *Canadian Journal of*
1897 *Earth Sciences*, v. 4, p. 271-298.
1898
- 1899 Gabrielse, H., Blusson, S.L., and Roddick, J.A., 1973, Flat River, Glacier Lake, and Wrigley
1900 Lake map-areas (95E, L, M), District of Mackenzie and Yukon Territory: *Geological*
1901 *Survey of Canada, Memoir 366*, 153 p.
1902
- 1903 Galán, E., and Pozo, M., 2011, Palygorskite and sepiolite deposits in continental environments:
1904 Description, genetic patterns, and sedimentary settings, *in* Galán, E., and Singer, A., editors,
1905 *Developments in Palygorskite-Sepiolite Research: A new outlook on these nanomaterials:*
1906 *Developments in Clay Science*, Amsterdam, Elsevier, p. 125-174.
1907
- 1908 Gat, J.R., and Bowser, C., 1991, The heavy isotope enrichment of water in coupled evaporative

- 1909 systems, *in* Taylor, H.P., O'Neil, J.R., Kaplan, I.R., editors, *Stable Isotope Geochemistry: A Tribute to Samuel Epstein: Geochemical Society Special Publication*, v. 3, p. 159–168.
1910
- 1911 Gawthorpe, R.L., Fraser, A.J., and Collier, R.E.L.L., 1994, Sequence stratigraphy in active
1912 extensional basins: implications for the interpretation of ancient basin-fills: *Marine and*
1913 *Petroleum Geology*, v. 11, p. 642–658.
- 1914 Gawthorpe, R.L., and Leeder, M.R., 2000, Tectono-sedimentary evolution of active extensional
1915 basins: *Basin Research*, v. 12, no. 3–4, p. 195–218, doi: 10.1111/j.1365-2117.2000.00121.x.
- 1916 Godderis, Y., Donnadieu, Y., Nedelec, A., Dupre, B., Dessert, C., Grard, A., Ramstein, G., and
1917 Francois, L., 2003, The Sturtian “snowball” glaciation: fire and ice: *Earth And Planetary*
1918 *Science Letters*, v. 211, no. 1–2, p. 1–12.
- 1919 Goodfellow, W., Cecile, M., and Leybourne, M., 1995, Geochemistry, petrogenesis, and tectonic
1920 wetting of Lower Paleozoic alkalic and potassic volcanic rocks, northern Canadian
1921 Cordilleran miogeocline: *Canadian Journal of Earth Sciences*, v. 32, no. 8, p. 1236–1254.
- 1922 Gordey, S.P., and Anderson, R.G., 1993, Evolution of the Cordilleran miogeocline, Nahanni map
1923 area (105I), Yukon and Northwest Territories: *Geological Survey of Canada, Memoir 428*,
1924 214 p.
- 1925 Gorjan, P., Veevers, J.J., and Walter, M.R., 2000, Neoproterozoic sulfur-isotope variation in
1926 Australia and global implications: *Precambrian Research*, v. 100, p. 151–179.
- 1927 Green, L.H., 1972, Geology of Nash Creek, Larsen Creek, and Dawson Creek map areas, Yukon
1928 Territory, *Geological Society of Canada, Memoir 364*, 157 p.
- 1929 Grotzinger, J.P., 1986, Evolution of Early Proterozoic passive margin carbonate platform,
1930 Rocknest Formation, Wopmay Orogen, Northwest Territories, Canada: *Journal of*
1931 *Sedimentary Petrology*, v. 56, no. 6, p. 831–847.
- 1932 Grotzinger, J.P., 1989, Facies and evolution of Precambrian carbonate depositional systems:
1933 Emergence of the modern platform archetype, *in* Crevello, P.D., Wilson, J.L., Sarg, J.F.,
1934 and Read, J.F., editors, *Controls on Carbonate Platform and Basin Development: Society*
1935 *for Sedimentary Geology (SEPM) Special Publication*, v. 44, p. 79–106.
- 1936 Grotzinger, J.P., and James, N.P., 2000, Precambrian carbonates: Evolution of understanding, *in*
1937 Grotzinger, J.P., and James, N.P., editors, *Carbonate Sedimentation and Diagenesis in the*
1938 *Evolving Precambrian World: Society for Sedimentary Geology (SEPM) Special*
1939 *Publication*, v. 67, p. 1–20.
- 1940 Halverson, G.P., 2006, A Neoproterozoic chronology, *in* Xiao, S., and Kaufman, A.J., editors,
1941 *Neoproterozoic Geobiology and Paleobiology: Topics in Geobiology*, Springer, New York,
1942 v. 27, p. 231–271.
- 1943 Halverson, G.P., and Hurtgen, M.T., 2007, Ediacaran growth of the marine sulfate reservoir:
1944 *Earth and Planetary Science Letters*, v. 263, p. 32–44.
- 1945 Halverson, G.P., Maloof, A.C., and Hoffman, P.F., 2004, The Marinoan glaciation
1946 (Neoproterozoic) in Svalbard: *Basin Research*, v. 16, p. 297–324.

- 1947 Halverson, G.P., Hoffman, P.F., Schrag, D.P., Maloof, A.C., and Rice, A., 2005, Toward a
1948 Neoproterozoic composite carbon isotope record: Geological Society of America Bulletin,
1949 v. 117, p. 1181–1207.
- 1950 Halverson, G.P., Wade, B.P., Hurtgen, M.T., and Barovich, K.M., 2010, Neoproterozoic
1951 Chemostratigraphy: Precambrian Research, v. 182, no. 4, p. 337–350.
- 1952 Halverson, G.P., Macdonald, F.A., Strauss, J.V., Smith, E.F., Cox, G.M., and Hubert-Theou, L.,
1953 2012, Updated definition and correlation of the lower Fifteenmile Group in the central and
1954 eastern Ogilvie Mountains, *in* MacFarlane, K.E., and Sack, P.J., editors, Yukon Exploration
1955 and Geology 2011: Yukon Geological Survey, p. 75–90.
- 1956 Halverson, G.P., Cox, G.C., Hurtgen, M.T., Sansjofre, P., Kunzmann, M., Strauss, J.V., and
1957 Macdonald, F.A., 2014, A continental flood basalt driver for Neoproterozoic climate and
1958 oxygenation: Geological Society of America Abstracts with Programs, p. 256–256.
- 1959 Hansen, V.L., Goodge, J.W., Keep, M., and Oliver, D.H., 1993, Asymmetric rift interpretation of
1960 the western North American margin: Geology, v. 21, p. 1067–1070.
- 1961 Harlan, S., Heaman, L., LeCheminant, A., and Premo, W., 2003, Gunbarrel mafic magmatic
1962 event: A key 780 Ma time marker for Rodinia plate reconstructions: Geology, v. 31, no. 12,
1963 p. 1053–1056.
- 1964 Hartnett, H.E., Keil, R.G., Hedges, J.I., Devol, A.H., 1998, Influence of oxygen exposure time on
1965 organic carbon preservation in continental margin sediments: Nature, v. 391, p. 572–574.
- 1966 Harwood, C.L., and Sumner, D.Y., 2011, Microbialites of the Neoproterozoic Beck Spring
Dolomite, Southern California: Sedimentology, v. 58, no. 6, p. 1648–1673, doi:
10.1111/j.1365-3091.2011.01228.x.
- 1972 Harwood, C.L., and Sumner, D.Y., 2012, Origins of microbial microstructures in the
1973 Neoproterozoic Beck Spring Dolomite: variations in microbial community and timing of
1974 lithification: Journal of Sedimentary Research, v. 82, no. 9, p. 709–722, doi:
1975 10.2110/jsr.2012.65.
- 1976 Heaman, L., LeCheminant, A., and Rainbird, R., 1992, Nature and timing of Franklin Igneous
1977 Events, Canada: implications for a Late Proterozoic mantle plume and the break-up of
1978 Laurentia: Earth And Planetary Science Letters, v. 109, p. 117–131.
- 1979 Hedges, J.I., Hu, F.S., Devol, A.H., Hartnett, H.E., Tsamakis, E., and Keil, R.G., 1999,
1980 Sedimentary organic matter preservation: an assessment and speculative synthesis:
1981 American Journal of Science, v. 299, p. 529–555.
- 1982 Helmstaedt, H., Eisbacher, G.H., and McGregor, J.A., 1979, Copper mineralization near an intra-
1983 Rapitan unconformity, Nite copper prospect, Mackenzie Mountains, Northwest Territories,
1984 Canada: Canadian Journal of Earth Sciences, v. 16, p. 50–59.
- 1985 Higgins, J.A., and Schrag, D.P., 2006, Beyond methane: Towards a theory for the Paleocene-
1986 Eocene Thermal Maximum: Earth and Planetary Science Letters, v. 245, p. 523–537.
- 1987 Hoffman, P.F., 1989, Precambrian geology and tectonic history of North America, *in* Bally, A.W.,

- 1988 and Palmer, A.R., editors, The geology of North America – an overview: Geological
1989 Society of America, The Geology of North America, v. A, p. 447-512.
- 1990 Hoffman, P.F., 1991, Did the breakout of Laurentia turn Gondwanaland inside-out?: Science, v.
1991 252, no. 5011, p. 1409–1412.
- 1992 Hoffman, P.F., Kaufman, A.J., Halverson, G.P., and Schrag, D.P., 1998, A Neoproterozoic
1993 snowball earth: Science, v. 281, p. 1342–1346.
- 1994 Hoffman, P.F., and Schrag, D.P., 2002, The snowball Earth hypothesis: testing the limits of
1995 global change: Terra Nova, v. 14, no. 3, p. 129–155.
- 1996 Hoffman, P.F., Halverson, G.P., Domack, E., Maloof, A.C., Swanson-Hysell, N., and Cox, G.M.,
1997 2012, Cryogenian glaciations on the southern tropical paleomargin of Laurentia (NE
1998 Svalbard and East Greenland), and a primary origin for the upper Russøya (Islay) carbon
1999 isotope excursion: Precambrian Research, v. 206-207, p. 137-158.
- 2000 Horodyski, R.J., and Knauth, P.L., 1994, Life on land in the Precambrian: Science, v. 263, no.
2001 5146, p. 494-498.
- 2002 Hover, V.C., Walter, L.M., Peacor, D.R., and Martini, A.M., 1999, Mg-smectite authigenesis in a
2003 marine evaporative environment, Salina Ometepe, Baja California: Clays and Clay
2004 Minerals, v. 47, no. 3, p. 252-268.
- 2005 Hughes, G.W., Abdine, S., and Girgis, M.H., 1992, Miocene biofacies development and
2006 geological history of the Gulf of Suez, Egypt: Marine and Petroleum Geology, v. 9, p. 2-28.
2007
- 2008 Hurtgen, M.T., Arthur, M.A., Suits, N.S., and Kaufman, A.J., 2002, the sulfur isotopic
2009 composition of Neoproterozoic seawater sulfate: Implications for a snowball Earth?: Earth
2010 and Planetary Science Letters, v. 203, p. 413-430.
- 2011 Jackson, J.A., and White, N.J., 1989, Normal faulting in the upper continental crust: observations
2012 from regions of active extension: Journal of Structural Geology, v. 11, p. 15-36.
2013
- 2014 Jackson, M.P.A., Warin, O.N., Woad, G.M., and Hudec, M.R., 2003, Neoproterozoic
2015 allochthonous salt tectonics during the Lufilian orogeny in the Katangan Copperbelt,
2016 central Africa: Geological Society of America Bulletin, v. 115, p. 314-330.
2017
- 2018 Jacobsen, S.B., and Kaufman, A.J., 1999, The Sr, C, and O isotopic evolution of Neoproterozoic
2019 seawater: Chemical Geology, v. 161, p. 37-57.
- 2020 Jaffrés, J.B.D., Shields, G.A., and Wallman, K., 2007, The oxygen isotope evolution of seawater:
2021 A critical review of a long-standing controversy and an improved geological water cycle
2022 model for the past 3.4 billion years: Earth Science Reviews, v. 83, p. 83-122.
- 2023 James, N., Narbonne, G., and Kyser, T., 2001, Late Neoproterozoic cap carbonates: Mackenzie
2024 Mountains, northwestern Canada: precipitation and global glacial meltdown: Canadian
2025 Journal of Earth Sciences, v. 38, no. 8, p. 1229–1262.
- 2026 Jefferson, C.W., ms, 1983, The Upper Proterozoic Redstone Copper Belt, Mackenzie Mountains,
2027 N.W.T.: Ph.D. thesis, The University of Western Ontario, London, 445 p.

- 2028 Jefferson, C.W., 1985, Uppermost Shaler Group and its contact with the Natkusiak basalts,
2029 Victoria Island, District of Franklin, Current Research Part A: Geological Survey of
2030 Canada, Paper 85-1A, p. 103-110.
- 2031 Jefferson, C.W., and Ruelle, J.C.L., 1986, The Late Proterozoic Redstone Copper Belt,
2032 Mackenzie Mountains, N.W.T., *in* Morin, J.A., editor, Mineral Deposits of the Northern
2033 Cordillera: Canadian Institute of Mining and Metallurgy, Special Volume 37, p. 154-168.
- 2034 Jefferson, C.W., and Parrish, R.R., 1989. Late Proterozoic stratigraphy, U–Pb zircon ages, and rift
2035 tectonics, Mackenzie Mountains, northwestern Canada: Canadian Journal of Earth Science,
2036 v. 26, p. 1784–1801.
- 2037 Jeletsky, J.A., 1962, Pre-Cretaceous Richardson Mountains Trough – its place in the tectonic
2038 framework of Arctic Canada and its bearing on some geosynclinal concepts: Transactions
2039 of the Royal Society of Canada, v. 56, p. 55-84.
2040
- 2041 Johnson, J.G., Murphy, M.A., 1984, Time-rock model for Siluro-Devonian continental shelf,
2042 western United States: Geological Society of America Bulletin, v. 95, p. 1349-1359.
- 2043 Johnson, J.G., Klapper, G., and Sandberg, C.A., 1985, Devonian eustatic fluctuations in
2044 Euramerica: Geological Society of America Bulletin, v. 96, p. 567-587.
- 2045 Jones, D.S., Maloof, A.C., Hurtgen, M.T., Rainbird, R.H., and Schrag, D.P., 2010, Regional and
2046 global chemostratigraphic correlation of the early Neoproterozoic Shaler Supergroup,
2047 Arctic Canada: Precambrian Research, v. 181, p. 43-63.
- 2048 Jones, D.S., Creel, R.C., Rios, B., and Santiago Ramos, D.P., 2015, Chemostratigraphy of an
2049 Ordovician–Silurian carbonate platform: $\delta^{13}\text{C}$ records below glacioeustatic exposure
2050 surfaces: Geology, v. 43, no. 1, p. 59-62.
- 2051 Kah, L.C., Sherman, A.G., Narbonne, G.M., Knoll, A.H., and Kaufman, A.J., 1999, $\delta^{13}\text{C}$
2052 stratigraphy of the Proterozoic Bylot Supergroup, Baffin Island, Canada: implications for
2053 regional lithostratigraphic correlation: Canadian Journal of Earth Science, v. 36, p. 313-332.
- 2054 Kah, L.C., 2000, Depositional $\delta^{18}\text{O}$ signatures in Proterozoic dolostones: constraints on seawater
2055 chemistry and early diagenesis, *in* Grotzinger, J.P., and James, N.P., editors, Carbonate
2056 Sedimentation and Diagenesis in the Evolving Precambrian World: Society for
2057 Sedimentary Geology (SEPM) Special Publication, v. 67, p. 345-360.
- 2058 Karlstrom, K.E., Bowring, S.A., Dehler, C.M., Knoll, A.H., Porter, S.M., Des Marais, D.J., Weil,
2059 A.B., Sharp, Z.D., Geissman, J.W., Elrick, M.B., Timmons, J.M., Crossey, L.J., and
2060 Davidek, K.L., 2000, Chuar Group of the Grand Canyon: Record of breakup of Rodinia,
2061 associated change in the global carbon cycle, and ecosystem expansion by 740 Ma:
2062 Geology, v. 28, no. 7, p. 619–622.
- 2063 Kaufman, A.J., Knoll, A.H., and Narbonne, G., 1997, Isotopes, ice ages, and terminal Proterozoic
2064 earth history: Proceedings of the National Academy of Sciences, v. 94, no. 13, p. 6600–
2065 6605.
- 2066 Kaufman, A.J., Williams, B.P., Johnston, D.T., Farquhar, J., Knoll, A.H., Butterfield, N.J., and
2067 Rainbird, R.H., 2007, Environmental change in the prelude to a Neoproterozoic ice age:

- 2068 Sulfur isotope evidence from the Shaler Supergroup, Northwest Territories, Canada:
2069 American Geophysical Union, Fall Meeting 2007, Abstract #V42A-01.
- 2070 Kendall, A.C., 2001, Late diagenetic calcitization of anhydrite from the Mississippian of
2071 Saskatchewan, western Canada: *Sedimentology*, v. 48, p. 29-55.
- 2072 Kendall, C.G.St.C., and Warren, J., 1987, A review of the origin and setting of tepees and their
2073 associated fabrics: *Sedimentology*, v. 34, p. 1007–1027.
- 2074
2075 Kingsbury-Stewart, E.M., Osterhout, S.L., Link, P.K., and Dehler, C.M., 2013, Sequence
2076 stratigraphy and formalization of the Middle Uinta Mountain Group (Neoproterozoic),
2077 central Uinta Mountains, Utah: A closer look at the western Laurentian Seaway at ca. 750
2078 Ma: *Precambrian Research*, v. 236, p. 65–84, doi: 10.1016/j.precamres.2013.06.015.
- 2079 Kirkham, A., 1997, Shoreline evolution, aeolian deflation and anhydrite distribution of the
2080 Holocene, Abu Dhabi: *GeoArabia*, v. 2, p. 403-416.
- 2081
2082 Kirshvink, J.L., 1992, Late Proterozoic low-latitude global glaciation: the snowball earth, *in* Klein,
2083 C., and Schopf, J.W., editors, *The Proterozoic Biosphere*: London, Cambridge University
2084 Press, p. 51-52.
- 2085
2086 Knauth, L.P., and Kennedy, M.J., 2009, The late Precambrian greening of the Earth: *Nature*, v.
2087 460, no. 7256, p. 728–732, doi: 10.1038/nature08213.
- 2088 Knoll, A.H., Hayes, J.M., Kaufman, A.J., Swett, K., and Lambert, I.B., 1986, Secular variation in
2089 carbon isotope ratios from Upper Proterozoic successions of Svalbard and East Greenland:
2090 *Nature*, v. 321, no. 6073, p. 832–838, doi: 10.1038/321832a0.
- 2091 Krapez, B., 1996, Sequence stratigraphic concepts applied to the identification of basin-filling
2092 rhythms in Precambrian successions: *Australian Journal of Earth Sciences*, v. 43, no. 4, p.
2093 355-380.
- 2094
2095 Kump, L.R., 1991, Interpreting carbon-isotope excursions: Strangelove oceans: *Geology*, v. 19, p.
2096 299-302.
- 2097
2098 Kump, L.R., Arthur, M.A., Patzkowsky, M.E., Gibbs, M.T., Pinkus, D.S., and Sheehan, P.M.,
2099 1999, A weathering hypothesis for glaciation at high atmospheric $p\text{CO}_2$ during the Late
2100 Ordovician: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 152, p. 173-187.
- 2101
2102 Kunzmann, M., Halverson, G.P., Macdonald, F.A., Hodgskiss, M., Sansjofre, P.D., Schumann, D.,
2103 and Rainbird, R.H., 2014, The early Neoproterozoic Chandindu Formation of the
2104 Fifteenmile Group in the Ogilvie Mountains, *in* MacFarlane, K.E., Nordling, M.G., and
2105 Sack, P.J., editors, *Yukon Exploration and Geology 2013*: Yukon Geological Survey, p. 93-
2106 107.
- 2107
2108 Land, L.S., 1983, The application of stable isotopes to studies of the origin of dolomite and to
2109 problems of diagenesis of clastic sediments, *in* Arthur, M.A., and Anderson, T.F., editors,
2110 *Stable Isotopes in Sedimentary Geology*: Society of Economic and Paleontologists
2111 Mineralogical (SEPM) Short Course No. 10, 4.1-4.22.
- 2112 LeCheminant, A.N., and Heaman, L.M., 1989, Mackenzie igneous events, Canada: Middle

- 2113 Proterozoic hotspot magmatism associated with ocean opening: *Earth And Planetary*
2114 *Science Letters*, v. 96, p. 38-48.
- 2115 Leeder, M.R., and Gawthorpe, R.L., 1987, Sedimentary models for extensional tilt-block/half-
2116 graben basins, *in* Coward, M.P., Dewey, J.F., and Hancock, P.L., editors, *Continental*
2117 *Extensional Tectonics: Geological Society of London Special Publication*, v. 28, p. 139-152.
2118
- 2119 Lenz, A.C., 1972, Ordovician to Devonian history of northern Yukon and adjacent District of
2120 Mackenzie: *Bulletin of Canadian Society of Petroleum Geologists*, v. 20, no. 2, p. 321-361.
2121
- 2122 Levy, M., and Christie-Blick, N., 1991, Tectonic Subsidence of the Early Paleozoic Passive
2123 Continental-Margin in Eastern California and Southern Nevada: *Geological Society of*
2124 *America Bulletin*, v. 103, no. 12, p. 1590–1606.
- 2125 Li, Z.-X., Bogdanova, S.V., Collins, A.S., Davidson, A., De Waele, B., Ernst, R.E., Fitzsimmons,
2126 I.C.W., Fuck, R.A., Gladkochub, D.P., Jacobs, J., Karlstrom, K.E., Lu, S., Natapov, L.M.,
2127 Pease, V., Pisarevsky, S.A., Thrane, K., Vernikovsky, V., 2008, Assembly, configuration,
2128 and break-up history of Rodinia: A synthesis: *Precambrian Research*, v. 160, no. 1-2, p.
2129 179–210, doi: 10.1016/j.precamres.2007.04.021.
- 2130 Li, Z.-X., Kinny, P.D., and Wang, J., 1999, The breakup of Rodinia: did it start with a mantle
2131 plume beneath South China?: *Earth And Planetary Science Letters*, v. 173, p. 171-181.
- 2132 Li, Z.-X., Evans, D.A.D., Halverson, G.P., 2013, Neoproterozoic glaciations in a revised global
2133 palaeogeography from the breakup of Rodinia to the assembly of Gondwanaland:
2134 *Sedimentary Geology*, v. 294, p. 219-232.
2135
- 2136 Lindsay, J.F., 2002, Supersequences, superbasins, supercontinents – evidence from the
2137 Neoproterozoic–Early Palaeozoic basins of central Australia: *Basin Research*, v. 14, p. 207-
2138 223.
2139
- 2140 Long, D.G.F., Rainbird, R.H., Turner, E.C., and MacNaughton, R.B., 2008, Early Neoproterozoic
2141 strata (Sequence B) of mainland northern Canada and Victoria and Banks Islands: a
2142 contribution to the Geological Atlas of the Northern Canadian Mainland Sedimentary
2143 Basin: *Geological Survey of Canada, Open File 5700*, 24 p.
2144
- 2145 Long, D.G.F., and Turner, E.C., 2012, Formal definition of the Neoproterozoic Mackenzie
2146 Mountains Supergroup (Northwest Territories), and formal stratigraphic nomenclature for
2147 terrigenous clastic units of the Katherine Group: *Geological Survey of Canada, Open File*
2148 *7112*, 40 p.
2149
- 2150 Lund, K., 2008, Geometry of the Neoproterozoic and Paleozoic rift margin of western Laurentia:
2151 implications for mineral deposit settings: *Geosphere*, v. 4, p. 429-444.
2152
- 2153 Lund, K., Aleinikoff, J.N., Evans, K.V., duBray, E.A., Dewitt, E.H., and Unruh, D.M., 2010,
2154 SHRIMP U-Pb dating of recurrent Cryogenian and Late Cambrian-Early Ordovician alkalic
2155 magmatism in central Idaho: Implications for Rodinian rift tectonics: *Bulletin of the*
2156 *Geological Society of America*, v. 122, no. 3-4, p. 430.
- 2157 Lloyd, M.R., 1964, Variations in the oxygen and carbon isotope ratios of Florida Bay mollusks
2158 and their environmental significance: *Journal of Geology*, v. 72, p. 84-111.

2159
2160 Macdonald, F.A., and Roots, C.F., 2010, Upper Fifteenmile Group in the Ogilvie Mountains and
2161 correlations of early Neoproterozoic strata in the northern Cordillera, *in* McFarlane, K.E.,
2162 Weston, L.H., and Blackburn, L.R., editors, Yukon Exploration and Geology 2009: Yukon
2163 Geological Survey, p. 237-252.
2164
2165 Macdonald, F.A., Schmitz, M.D., Crowley, J.L., Roots, C.F., Jones, D.S., Maloof, A.C., Strauss,
2166 J.V., Cohen, P.A., Johnston, D.T., and Schrag, D.P., 2010, Calibrating the Cryogenian:
2167 Science, v. 327, no. 5970, p. 1241–1243, doi: 10.1126/science.1183325.

2168 Macdonald, F.A., Smith, E.F., Strauss, J.V., Cox, G.M, Halverson, G.P. and Roots, C.F., 2011,
2169 Neoproterozoic and early Paleozoic correlations in the western Ogilvie Mountains, Yukon,
2170 *in* MacFarlane, K.E., Weston, L.H., and Relf C., editors, Yukon Exploration and Geology
2171 2010: Yukon Geological Survey, p. 161-182

2172 Macdonald, F.A., Halverson, G.P., Strauss, J.V., Smith, E.F., Cox, G.M., Sperling, E.A., and
2173 Roots, C.F., 2012, Early Neoproterozoic basin formation in Yukon, Canada: Implications
2174 for the make-up and break-up of Rodinia: Geoscience Canada, v. 39, p. 77-99.
2175

2176 Macdonald, F.A., Strauss, J.V., Sperling, E.A., Halverson, G.P., Narbonne, G.M., Johnston, D.T.,
2177 Kunzmann, M., Schrag, D.P., and Higgins, J.A., 2013a, The stratigraphic relationship
2178 between the Shuram carbon isotope excursion, the oxygenation of Neoproterozoic oceans,
2179 and the first appearance of the Ediacara biota and bilaterian trace fossils in northwestern
2180 Canada: Chemical Geology, v. 362, p. 250-272, doi: 10.1016/j.chemgeo.2013.05.032.

2181 Macdonald, F., Prave, A., Petterson, R., Smith, E., Pruss, S., Oates, K., Waechter, F., Trotzuk, D.,
2182 and Fallick, A., 2013b, The Laurentian record of Neoproterozoic glaciation, tectonism, and
2183 eukaryotic evolution in Death Valley, California: Geological Society of America Bulletin, v.
2184 125, no. 7-8, p. 1203–1223.

2185 Mahon, R.C., Dehler, C.M., Link, P.K., Karlstrom, K.E., and Gehrels, G.E., 2014,
2186 Geochronologic and stratigraphic constraints on the Mesoproterozoic and Neoproterozoic
2187 Pahrump Group, Death Valley, California: A record of the assembly, stability, and breakup
2188 of Rodinia: Geological Society of America Bulletin, v. 126, no. 5-6, p. 652–664, doi:
2189 10.1130/B30956.1.

2190 Mann, P., Hempton, M.R., Bradley, D.C., and Burke, K., 1983, Development of pull-apart basins:
2191 Journal of Geology, v. 91, p. 529-554.
2192

2193 McCay, G.A., Prave, A.R., Alsop, G.I., Fallick, A.E., 2006, Glacial trinity: Neoproterozoic Earth
2194 history within the British-Irish Caledonides: Geology, v. 34, p. 909-912.

2195 McDonough, M.R., and Parrish, R.R., 1991, Proterozoic gneisses of the Malton Complex, near
2196 Valemount, British Columbia: U-Pb ages and Nd isotopic signatures: Canadian Journal of
2197 Earth Sciences, v. 28, p. 1202-1216.
2198

2199 McKenzie, J.A., 1981, Holocene dolomitization of calcium carbonate sediments from the coastal
2200 sabkhas of Abu Dhabi, UAE: a stable isotope study: Journal of Geology, v. 89, p. 185–198.

2201 Medig, K.P., Thorkelson, D.J., and Dunlop, R.L., 2010, The Proterozoic Pinguicula Group:
2202 stratigraphy, contact relationships, and possible correlations, *in* MacFarlane, K.E., Weston,

- 2203 L.H., and Blackburn, L.R., editors, Yukon Exploration and Geology 2009: Yukon
2204 Geological Survey, p. 265-278.
- 2205 Medig, K.P.R., Thorkelson, D.J., Davis, W.J., Rainbird, R.H., Gibson, H.D., Turner, E.C., and
2206 Marshall, D.D., 2014, Pinning northeastern Australia to northwestern Laurentia in the
2207 Mesoproterozoic: Precambrian Research, v. 249, p. 88–99, doi:
2208 10.1016/j.precamres.2014.04.018.
- 2209 Mercier, E., 1989, Evenements tectoniques d'origine compressive dans le Protérozoïque du nord
2210 de la Cordillere canadienne (montagnes Ogilvie, Yukon): Canadian Journal of Earth
2211 Sciences, v. 26, p. 199-205.
- 2212 Milliken, K.L., 1979, The silicified evaporite syndrome – two aspects of silicification history of
2213 former evaporite nodules from southern Kentucky and northern Tennessee: Journal of
2214 Sedimentary Research, v. 49, p. 245–256. doi: 10.1306/212F7707-2B24-11D7-
2215 8648000102C1865D.
- 2216 Millot, G., 1970, Geology of Clays: London, Springer–Verlag, Chapman and Hall, 429 p.
- 2217 Milton, J.E., Hickey, K.A., and Gleeson, S.A., 2014, New U-Pb constraints on Gunbarrel
2218 volcanism and the break-up of Rodinia: The 775 Ma Little Dal basalts: Geological Society
2219 of America Abstracts with Programs, v. 46, n. 6, p. 828.
- 2220 Monger, J., and Price, R., 2002, The Canadian Cordillera: Geology and Tectonic Evolution:
2221 Canadian Society for Economic Geologists Recorder, p. 17-36.
- 2222 Montenat, C., Orszag-Sperber, F., Plaziat, J.-C., and Purser, B.H., 1998, The sedimentary record
2223 of the initial stages of Oligo-Miocene rifting in the Gulf of Suez and the northern Red Sea,
2224 *in* Purser, B.H., and Bosence, D.W.J., editors, Sedimentation and Tectonics of Rift Basins:
2225 Red Sea–Gulf of Aden: Chapman and Hall, London, p. 146-xxx.
2226
- 2227 Moodley, L., Middelburg, J.J., Herman, P.M.J., and de Lange, G.J., 2005, Oxygenation and
2228 organic-matter preservation in marine sediments: direct experimental evidence from ancient
2229 organic carbon-rich deposits: Geology, v. 33, p. 889-892.
2230
- 2231 Morgan, W.J., 1981, Hot spot tracks and the opening of the Atlantic and Indian Oceans: The
2232 oceanic lithosphere, *in* Emiliani, C., editor, The Sea – Ideas and Observations on Progress
2233 in the Study of the Seas: New York, Wiley Interscience, v. 7, p. 443-487.
- 2234 Morrow, D.W., 1999, Lower Paleozoic stratigraphy of northern Yukon Territory and
2235 northwestern District of Mackenzie: Geological Survey of Canada, Bulletin 538, 202 p.
2236
- 2237 Moynihan, D., 2014, Bedrock geology of NTS 106B/04, Eastern Rackla Belt, *in* MacFarlane,
2238 K.E., Nordling, M.G., and Sack, P.J., editors, Yukon Exploration and Geology 2013:
2239 Yukon Geological Survey, p. 147-167.
2240
- 2241 Mustard, P.S., ms, 1990, Sedimentology and Tectonic Evolution of the Upper Proterozoic Mount
2242 Harper Group Sedimentary Rocks, Yukon Territory: Ph.D. thesis, Carleton University,
2243 Ottawa, 330 p.
- 2244 Mustard, P.S., 1991, Normal faulting and alluvial fan deposition, basal Windermere Tectonic

- 2245 Assemblage, Yukon, Canada: Geological Society of America Bulletin, v. 103, no. 10, p.
2246 1346–1364.
- 2247 Mustard, P.S., and Donaldson, J.A., 1990, Paleokarst breccias, calcretes, silcretes and fault talus
2248 breccias at the base of Upper Proterozoic “Windermere” strata, northern Canadian
2249 Cordillera: *Journal of Sedimentary Petrology*, v. 60, no. 4, p. 525–539.
- 2250 Mustard, P.S., and Roots, C.F., 1997, Rift-related volcanism, sedimentation, and tectonic setting
2251 of the Mount Harper Group, Ogilvie Mountains, Yukon Territory: Geological Survey of
2252 Canada, Bulletin 492, 92 p.
2253
- 2254 Naehr, T.H., Eichhubl, P., Orphan, V.J., Hovland, M., Paull, C.K., Ussler III, W., Lorenson, T.D.,
2255 and Greene, H.G., 2007, Authigenic carbonate formation at hydrocarbon seeps in
2256 continental margin sediments: A comparative study: *Deep Sea Research Part II: Tropical*
2257 *Studies in Oceanography*, v. 54, no. 11-13, p. 1268-1291.
- 2258 Narbonne, G.M., and Aitken, J., 1995, Neoproterozoic of the Mackenzie Mountains,
2259 Northwestern Canada: *Precambrian Research*, v. 73, p. 101–121.
- 2260 Narbonne, G.M., Kaufman, A.J., and Knoll, A.H., 1994, Integrated chemostratigraphy and
2261 biostratigraphy of the Windermere Supergroup, northwestern Canada: Implications for
2262 Neoproterozoic correlations and the early evolution of animals: *Geological Society of*
2263 *America Bulletin*, v. 106, p. 1281–1292.
- 2264 Norris, D.K., 1982, Snake River: Geological Survey of Canada, Map 1529A, scale 1:250,000.
2265
- 2266 Norris, D.K., 1997, Geology and mineral and hydrocarbon potential of northern Yukon Territory
2267 and northwestern District of Mackenzie: Geological Survey of Canada, Bulletin 422, 397 p.
2268
- 2269 Ootes, L., Sandeman, H., Lemieux, Y., and Leslie, C., 2008, The 780 Ma Tsezotene sills,
2270 Mackenzie Mountains: a field, petrographical, and geochemical study: Northwest
2271 Territories Geoscience Office, NWT Open File Report 2008-011, 21 p.
2272
- 2273 Orszag-Sperber, F., Harwood, G., Kendall, A., and Purser, B.H., 1998, A review of the evaporites
2274 of the Red Sea-Gulf of Suez rift, *in* Purser, B.H., and Bosence, D.W.J., editors,
2275 *Sedimentation and Tectonics of Rift Basins: Red Sea–Gulf of Aden*: London, Chapman and
2276 Hall, p. 409-426.
2277
- 2278 Park, J.K., Roots, C.F., and Brunet, N., 1992, Paleomagnetic evidence for rotation in the
2279 Neoproterozoic Mount Harper volcanic complex, Ogilvie Mountains, Yukon Territory:
2280 *Current Research, Part E. Geological Survey of Canada Paper*, p. 1–10.
- 2281 Park, J., Buchan, K., and Harlan, S., 1995, A proposed giant radiating dyke swarm fragmented by
2282 the separation of Laurentia and Australia based on paleomagnetism of ca. 780 Ma mafic
2283 intrusions in western North America: *Earth And Planetary Science Letters*, v. 132, p. 129-
2284 139.
- 2285 Parrish, R.R., and Scammell, R.J., 1988, The age of the Mount Copeland syenite gneiss and its
2286 metamorphic zircons, Monashee Complex, southeastern British Columbia: radiogenic age
2287 and isotopic studies, report 2: Geological Survey of Canada, Paper 88-2, p. 21-28.
2288

- 2289 Patterson, W.P., and Walter, L.M., 1994, Depletion of ^{13}C in seawater ΣCO_2 on modern
2290 carbonate platforms: Significance for carbon isotopic record of carbonates: *Geology*, v. 22,
2291 p. 885-888.
2292
- 2293 Pierre, C., and Rouchy, J.M., 1988, Carbonate replacements after sulphate evaporites in the
2294 Middle Miocene of Egypt: *Journal of Sedimentary Petrology*, v. 58, p. 446-456.
- 2295 Pope, M.C., and Grotzinger, J.P., 2003, Paleoproterozoic Stark Formation, Athapuscow Basin,
2296 Northwest Canada: Record of cratonic-scale salinity crisis: *Journal of Sedimentary*
2297 *Research*, v. 73, no. 2, p. 280-295.
- 2298 Pope, M.C., Grotzinger, J.P., and Schreiber, B.C., 2000, Evaporitic subtidal stromatolites
2299 produced by *in situ* precipitation: textures, facies associations, and temporal significance:
2300 *Journal of Sedimentary Research*, v. 70, no. 5, p. 1139-1151.
- 2301 Porter, S., and Knoll, A., 2000, Testate amoebae in the Neoproterozoic Era: evidence from vase-
2302 shaped microfossils in the Chuar Group, Grand Canyon: *Paleobiology*, v. 26, no. 3, p. 360-
2303 385.
- 2304 Posamentier, H.W., and Allen, G.P., 1999, Siliciclastic sequence stratigraphy: concepts and
2305 applications: *Society of Economic Paleontologists and Mineralogists, Concepts in*
2306 *Sedimentology and Paleontology Number 7*, 210 p.
- 2307 Pratt, B., 2010, Peritidal Carbonates, *in* James, N.P., and Dalrymple, R.W., editors, *Facies*
2308 *Models 4: Geological Association of Canada GEOText 6*, p. 401-420.
- 2309 Prave, A., 1999, Two diamictites, two cap carbonates, two $\delta^{13}\text{C}$ excursions, two rifts: The
2310 Neoproterozoic Kingston Peak Formation, Death Valley, California: *Geology*, v. 27, no. 4,
2311 p. 339-342.
- 2312 Prave, A., Fallick, A., Thomas, C.W., and Graham, C.M., 2009, A composite C-isotope profile
2313 for the Neoproterozoic Dalradian Supergroup of Scotland and Ireland: *Journal Of The*
2314 *Geological Society of London*, v. 166, p. 129-135.
- 2315 Prince, J.K.G., ms, 2014, Sequence stratigraphic, lithostratigraphic, and stable isotope analysis of
2316 the Minto Inlet Formation and Kilian Formation of the Shaler Supergroup, Northwest
2317 Territories: MSc Thesis, Carleton University, 139 p.
2318
- 2319 Prokoph, A., Shields, G.A., and Veizer, J., 2008, Compilation and time-series analysis of a
2320 marine carbonate $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, $^{87}\text{Sr}/^{86}\text{Sr}$ and $\delta^{34}\text{S}$ database through Earth history: *Earth-*
2321 *Science Reviews*, v. 87, p. 113-133.
2322
- 2323 Purser, B.H., 1973, *The Persian Gulf: Holocene Carbonate Sedimentation and Diagenesis in a*
2324 *Shallow Epicontinental Sea*: Berlin, Springer-Verlag, 471 p.
2325
- 2326 Pyle, L.J., and Barnes, C.R., 2003, Lower Paleozoic stratigraphic and biostratigraphic correlations
2327 in the Canadian Cordillera: implications for the tectonic evolution of the Laurentian
2328 margin: *Canadian Journal of Earth Sciences*, v. 40, no. 12, p. 1739-1753.
2329

- 2330 Rainbird, R.H., ms, 1991, Stratigraphy, sedimentology, and tectonic setting of the upper Shaler
2331 Group, Victoria Island, Northwest Territories: Ph.D. thesis, University of Western Ontario,
2332 London, 257 p.
2333
- 2334 Rainbird, R.H., 1992, Anatomy of a large-scale braidplain quartz-arenite from the Neoproterozoic
2335 Shaler Group, Victoria Island, NWT, Canada: *Canadian Journal of Earth Sciences*, v. 29, p.
2336 2537-2550.
2337
- 2338 Rainbird, R.H., 1993, The sedimentary record of mantle plume uplift preceding eruption of the
2339 Neoproterozoic Natkusiak flood basalt: *The Journal of Geology*, v. 101, no. 3, p. 305-318.
- 2340 Rayner, N., and Rainbird, R.H., 2013, U-Pb geochronology of the Shaler Supergroup, Victoria
2341 Island, northwest Canada: 2009-2013: Geological Survey of Canada, Open File 7419, 62 p.
2342
- 2343 Rainbird, R., Jefferson, C., and Young, G., 1996, The early Neoproterozoic sedimentary
2344 succession B of northwestern Laurentia: Correlations and paleogeographic significance:
2345 *Geological Society of America Bulletin*, v. 108, no. 4, p. 454-470.
- 2346 Rainbird, R., Jefferson, C.W., Halverson, G.P., and Thomson, D., 2014, Features of Rodinia
2347 break-up in the Minto Inlier with comparisons to the northern Cordillera: *Geological*
2348 *Society of America Abstracts with Programs*, v. 46, n. 6, p. 828.
- 2349 Read, J.F., 1982, Carbonate platforms of passive (extensional) continental margins: Types,
2350 characteristics and evolution: *Tectonophysics*, v. 81, no. 3-4, p. 195-212.
- 2351 Reading, H.G., 1980, Characteristics and recognition of strike-slip fault systems, *in* Balance, P.F.,
2352 and Reading, H.G., editors, *Sedimentation in Oblique-Slip Mobile Zones*: International
2353 Association of Sedimentologists Special Publication No. 4, p. 7-26.
2354
- 2355 Reimers, C.E., Ruttenberg, K.C., Canfield, D.E., Christiansen, M.B., and Martin, J.B., 1996,
2356 Porewater pH and authigenic phases formed in the uppermost sediments of the Santa
2357 Barbara Basin: *Geochimica et Cosmochimica Acta*, v. 60, p. 4037-4057.
- 2358 Riboulleau, A., Baudin, F., Deconinck, J.-F., Derenne, S., Largeau, C., and Tribouvillard, N., 2003,
2359 Depositional conditions and organic matter preservation pathways in an epicontinental
2360 environment: the Upper Jurassic Kashpir Oil Shales (Volga Basin, Russia):
2361 *Palaeogeography, Palaeoclimatology, and Palaeoecology*, p. 1-27.
- 2362 Richardson, M., and Arthur, M.A., 1988, The Gulf of Suez-northern Red Sea Neogene rift: a
2363 quantitative basin analysis: *Marine and Petroleum Geology*, v. 5, p. 247-270.
- 2364 Rooney, A.D., Macdonald, F.A., Strauss, J.V., Dudás, F.O., Hallmann, C., and Selby, D., 2014,
2365 Re-Os geochronology and coupled Os-Sr isotope constraints on the Sturtian snowball:
2366 *Proceedings of the National Academy of Sciences*, v. 111, p. 51-56.
- 2367 Rooney, A.D., Strauss, J.V., Brandon, A.D., and Macdonald, F.A., 2015, A Cryogenian
2368 chronology: Two long-lasting, synchronous Neoproterozoic glaciations: *Geology*, v. 43, n.
2369 5, p. 459-462.
2370

- 2371 Roots, C.F., ms, 1987, Regional tectonic setting and evolution of the Late Proterozoic Mount
2372 Harper volcanic complex, Ogilvie Mountains, Yukon: Ph.D. thesis, Carleton University,
2373 Ottawa, 180 p.
2374
- 2375 Roots, C.F., and Thompson, R.I., 1992, Long-lived basement weak zones and their role in
2376 extensional magmatism in the Ogilvie Mountains, Yukon Territory, *in* Bartholomew, M.J.,
2377 Hyndman, D.W., Mogk, D.W., and Mason, R., editors, Basement Tectonics and
2378 Characterization of Ancient and Mesozoic Continental Margins – Proceedings of the 8th
2379 International Conference in Basement Tectonics: Dordrecht, Kluwer Academic Publishers,
2380 p. 359-372.
2381
- 2382 Ross, G., 1991, Tectonic Setting of the Windermere Supergroup Revisited: *Geology*, v. 19, no. 11,
2383 p. 1125–1128.
- 2384 Sandeman, H.A., Ootes, L., Cousens, B., and Kilian, T., 2014, Petrogenesis of Gunbarrel
2385 magmatic rocks: Homogeneous continental tholeiites associated with extension and rifting
2386 of Neoproterozoic Laurentia: *Precambrian Research*, v. 252, p. 166-179.
2387
- 2388 Saunders, A.D., Jones, S.M., Morgan, L.A., Pierce, K.L., Widdowson, M., and Xu, Y.G., 2007,
2389 Regional uplift associated with continental large igneous provinces: The roles of mantle
2390 plumes and the lithosphere: *Chemical Geology*, v. 241, p. 282-318.
- 2391 Schrag, D.P., Berner, R.A., Hoffman, P.F., and Halverson, G.P., 2002, On the initiation of a
2392 snowball Earth: *Geochemistry, Geophysics, Geosystems*, v. 3, no. 6, p. 1-21.
- 2393 Schrag, D.P., Higgins, J.A., Macdonald, F.A., and Johnston, D.T., 2013, Authigenic carbonate
2394 and the history of the global carbon cycle: *Science*, v. 339, no. 6119, p. 540–543, doi:
2395 10.1126/science.1229578.
- 2396 Schreiber, B.C., Tucker, M.E., and Till, R., 1986, Arid shorelines and evaporites, *in* Reading,
2397 H.G., editor, *Sedimentary Environments and Facies*: Oxford, Blackwell, p. 189–228.
2398
- 2399 Schreiber, B.C., and El Tabakh, M., 2000, Deposition and early alteration of evaporites:
2400 *Sedimentology*, v. 47, p. 215-238.
- 2401 Schwab, D.L., Thorkelson, D.J., Mortensen, J.K., Creaser, R.A., and Abbott, J.G., 2004, The Bear
2402 River dykes (1265–1269 Ma): westward continuations of the Mackenzie dyke swarm into
2403 Yukon, Canada: *Precambrian Research*, v. 133, p. 175-186.
2404
- 2405 Selley, D., Broughton, D., Scott, R.J., Hizman, M., Bull, S.W., Large, R.R., McGoldrick, P.J.,
2406 Croaker, M and Pollington, N., 2005, A new look at the geology of the Zambian
2407 Copperbelt: *Society of Economic Geologists*, v. 100, p. 965-1000
2408
- 2409 Sharp, I.R., Gawthorpe, R.L., Armstrong, B., and Underhill, J.R., 2000, Fault-propagation folding
2410 in extensional settings: Examples of structural style and syn-rift sedimentary response from
2411 the Suez Rift, Sinai, Egypt: *Geological Society of America Bulletin*, v. 112, p. 1877-1899.
2412
- 2413 Shearman, D.J., 1978, Evaporites of coastal sabkhas, *in* Dean, W.E., and Schreiber, B.C., editors,
2414 *Marine Evaporites*: Society of Economic Paleontologists and Mineralogists (SEPM) Short
2415 Course 4, p. 6–42.
2416

2417 Slingerland, R., Kump, L.R., Arthur, M.A., Fawcett, P.J., Sageman, B.B., and Barron, E.J., 1996,
2418 Estuarine circulation in the Turonian Western Interior seaway of North America:
2419 Geological Society of America Bulletin, v. 108, p. 941-952.
2420

2421 Smith, E.F., Macdonald, F.A., Crowley, J.L., Hodgin, E.B., and Schrag, D.P., 2015, Tectono-
2422 stratigraphic evolution of the c. 780-730 Ma Beck Spring Dolomite: Basin formation in the
2423 core of Rodinia, *in* Li, Z.X., Evans, D.A.D., and Murphy, J.B., editors, Supercontinent
2424 Cycles Through Earth History: Geological Society of London, Special Publications, v. 424,
2425 p. xx.
2426

2427 Stewart, J.H., 1972, Initial deposits in the Cordilleran Geosyncline: Evidence of a late
2428 Precambrian (<850 m.y.) continental separation: Geological Society of America Bulletin, v.
2429 83, no. 5, p. 1345-1360.

2430 Stoessell, R.K., and Hay, R.L., 1978, The geochemical origin of sepiolite and kerolite at
2431 Amboseli, Kenya: Contributions to Mineralogy and Petrology, v. 65, no. 3, p. 255–267.

2432 Strauss, J.V., Rooney, A.D., Macdonald, F.A., Brandon, A.D., and Knoll, A.H., 2014a, 740 Ma
2433 vase-shaped microfossils from Yukon, Canada: Implications for Neoproterozoic
2434 chronology and biostratigraphy: *Geology*, v. 42, no. 8, p. 659-662.
2435

2436 Strauss, J.V., Roots, C.F., Macdonald, F.A., Halverson, G.P., Eyser, A., and Colpron, M., 2014b,
2437 Geological map of the Coal Creek inlier, Ogilvie Mountains (NTS 116B/10-15 and 116C/9,
2438 16): Yukon Geological Survey, Open File 2014-15, 1:100,000.

2439 Sun, X., and Turchyn, A.V., 2014, Significant contribution of authigenic carbonate to marine
2440 carbon burial: *Nature Geoscience*, v. 7, p. 201-204.

2441 Swart, P.K., and Kennedy, M.J., 2012, Does the global stratigraphic reproducibility of $\delta^{13}\text{C}$ in
2442 Neoproterozoic carbonates require a marine origin? A Pliocene-Pleistocene comparison:
2443 *Geology*, v. 40, no. 1, p. 87–90.

2444 Swett, K., and Knoll, A.H., 1989, Marine pisolites from Upper Proterozoic carbonates of East
2445 Greenland and Spitsbergen: *Sedimentology*, v. 36, p. 75-93.
2446

2447 Thomas, W.A., 1990, Controls on the location of transverse zones in thrust belts: *Eclogae*
2448 *Geologicae Helvetiae*, v. 83, p. 727-744.
2449

2450 Thomson, D., Rainbird, R.H., and Dix, G., 2014, Architecture of a Neoproterozoic intracratonic
2451 carbonate ramp succession: Wynnatt Formation, Amundsen Basin, Arctic Canada:
2452 *Sedimentary Geology*, v. 299, p. 119-138.
2453

2454 Thomson, D., Rainbird, R.H., and Krapez, B., 2015a, Sequence and tectonostratigraphy of the
2455 Neoproterozoic (Tonian-Cryogenian) Amundsen Basin prior to supercontinent (Rodinia)
2456 break-up: *Precambrian Research*, v. 263, p. 246–259.
2457

2458 Thomson, D., Rainbird, R.H., Planavsky, N., Lyons, T.W., and Bekker, A., 2015b,
2459 Chemostratigraphy of the Shaler Supergroup, Victoria Island, NW Canada: A record of
2460 ocean composition prior to the Cryogenian glaciations: *Precambrian Research*, v. 263, p.
2461 232–245.
2462

- 2463 Thompson, R.I., and Roots, C.F., 1992, Ogilvie Mountains project, Yukon: A new regional
2464 mapping program: Current Research, Part A, Geological Survey of Canada, Paper 82-1A, p.
2465 403-411.
- 2466
- 2467 Thompson, R.I., Mercier, B., and Roots, C.F., 1987, Extension and its influence on Canadian
2468 Cordilleran passive-margin evolution, *in* Coward, M.P., Dewey, J.F., and Hancock, P.L.,
2469 editors, *Continental Extensional Tectonics*: London, Geological Society Special Publication,
2470 v. 28, p. 409-417.
- 2471 Thompson, R.I., Roots, C.F., and Mustard, P.S., 1994, Geology of Dawson map area (116B, C,
2472 northeast of Tintina Trench): Geological Survey of Canada, Open File 2849.
- 2473
- 2474 Thorkelson, D.J., 2000, Geology and mineral occurrences of the Slat Creek, Fairchild Lake, and
2475 “Dolores Creek” areas, Wernecke Mountains, Yukon Territory (106D/16, 106C/13,
2476 106C/14): Exploration and Geological Services Division, Yukon Region, Indian and
2477 Northern Affairs Canada, Bulletin 10, 80 p.
- 2478 Thorkelson, D., Abbott, J., Mortensen, J., Creaser, R., Villeneuve, M., McNicoll, V., and Layer,
2479 P., 2005, Early and Middle Proterozoic evolution of Yukon, Canada: *Canadian Journal of*
2480 *Earth Sciences*, v. 42, no. 6, p. 1045–1071, doi: 10.1139/E05-075.
- 2481 Thorsteinsson, R., and Tozer, E.T., 1962, Banks, Victoria, and Sefansson Islands, Arctic
2482 Archipelago: Geological Survey of Canada, Memoir 330, 85 p.
- 2483
- 2484 Tosca, N.J., 2015, Geochemical pathways for Mg-silicate formation, *in* Pozo, M., ed., *Magnesian*
2485 *Clays: Characterization, origins and applications*: Association Internationale pour l’Etude
2486 *des Argiles* Special Publication, v. 2, 1-48.
- 2487
- 2488 Tosca, N.J. and Wright, V.P., 2015, Diagenetic pathways linked to labile Mg-clays in lacustrine
2489 carbonate reservoirs: A model for the origin of secondary porosity in the Cretaceous Pre-
2490 Salt Barra Velha Formation, offshore Brazil: Geological Society of London, Special
2491 Publication xxx, p. xxx.
- 2492
- 2493 Tosca, N.J., Macdonald, F.A., Strauss, J.V., Johnston, D.T., and Knoll, A.H., 2011, Sedimentary
2494 talc in Neoproterozoic carbonate successions: *Earth And Planetary Science Letters*, v. 306,
2495 no. 1-2, p. 11–22, doi: 10.1016/j.epsl.2011.03.041.
- 2496 Tucker, M.E., 1982, Precambrian dolomites: petrographic and isotopic evidence that they differ
2497 from Phanerozoic dolomites: *Geology*, v. 10, p. 7–12.
- 2498 Tucker, M.E., 1985, Shallow-marine carbonate facies and facies models, *in* Brenchley, P.J., and
2499 Williams, B.P.J., editors, *Sedimentology: Recent Developments and Applied Aspects*:
2500 *Special Publications of the Geological Society of London*, v. 18, p. 139–161.
- 2501
- 2502 Turner, E.C., 2011, Stratigraphy of the Mackenzie Mountains Supergroup in the Wernecke
2503 Mountains, Yukon, *in* MacFarlane, K.E., Weston, L.H., and Relf, C., editors, *Yukon*
2504 *Exploration and Geology 2010*: Yukon Geological Survey, p. 207-231.
- 2505 Turner, E.C., Roots, C.F., MacNaughton, R.B., Long, D.G.F., Fischer, B.J., Gordey, S.P., Martel,
2506 E., and Pope, M.C., 2011, Chapter Three: Stratigraphy, *in* Martel, E., Turner, E.C., and
2507 Fischer, B.J., editors, *Geology of the central Mackenzie Mountains of the northern Canadian*

- 2508 Cordillera, Sewki Mountain (105P), Mount Eduni (106A), and northwestern Wrigley Lake
2509 (95M) map-areas, Northwest Territories, NWT Special Volume 1, NWT Geoscience Office,
2510 p. 31-193.
2511
- 2512 Turner, E.C., and Long, D.G.F., 2008, Basin architecture and syndepositional fault activity during
2513 deposition of the Neoproterozoic Mackenzie Mountains supergroup, Northwest Territories,
2514 Canada: Canadian Journal of Earth Sciences, v. 45, p. 1159-1184.
- 2515 Turner, E.C., and Long, D.G.F., 2012, Formal definition of the Neoproterozoic Mackenzie
2516 Mountains Supergroup (Northwest Territories), and formal stratigraphic nomenclature for
2517 its carbonate and evaporite formations: Geological Survey of Canada, Open File 7112, 57 p.
- 2518 Tziperman, E., Halevy, I., Johnston, D.T., Knoll, A.H., and Schrag, D.P., 2011, Biologically
2519 induced initiation of Neoproterozoic snowball-Earth events: Proceedings of the National
2520 Academy of Sciences, v. 108, no. 37, p. 15091–15096.
- 2521 Ulmer-Scholle, D.S., and Scholle, P.A., 1994, Replacement of evaporites within the Permian Park
2522 City Formation, Bighorn Basin, Wyoming, USA: Sedimentology, v. 41, p. 1203-1222.
- 2523 Ulmer-Scholle, D.S., Scholle, P.A., and Brady, P.V., 1993, Silicification of evaporites in Permian
2524 (Guadalupian) back-reef carbonates of the Delaware Basin, West Texas and New Mexico:
2525 Journal of Sedimentary Research, v. 63, p. 955-965.
- 2526 van Acken, D., Thomson, D., Rainbird, R.H., and Creaser, R.A., 2013, Constraining the
2527 depositional history of the Neoproterozoic Shaler Supergroup, Amundsen Basin, NW
2528 Canada: Rhenium-osmium dating of black shales from the Wynniatt and Boot Inlet
2529 formations: Precambrian Research, v. 236, p. 124–131.
- 2530 van Straaten, L.M.J.U., 1954, Composition and structure of Recent marine sediments in the
2531 Netherlands: Leidsche Geologische Mededeelingen, v. 19, p. 1-110.
2532
- 2533 Villeneuve, M.E., Theriault, R.J., and Ross, G.M., 1991, U-Pb ages and Sm-Nd signature of two
2534 subsurface granites from the Fort Simpson magnetic high, northwest Canada: Canadian
2535 Journal of Earth Science, v. 28, p. 1003-1008.
2536
- 2537 Vogl, J.J., Min, K., Carmenate, A., Foster, D.A., and Marsellos, A., 2014, Miocene regional
2538 hotspot-related uplift, exhumation, and extension north of the Snake River Plain: Evidence
2539 from apatite (U-Th)/He thermochronology: Lithosphere, v. 6, no. 2, p. 108-123.
- 2540 Walter, M.R., Veevers, J.J., Calver, C.R., and Grey, K., 1995, Neoproterozoic stratigraphy of the
2541 Centralian Superbasin, Australia: Precambrian Research, v. 73, p. 173–195.
- 2542 Warren, J.K., 1989, Evaporite Sedimentology: Importance in Hydrocarbon Accumulation:
2543 Prentice Hall, Englewood Cliffs, N.J., 285 p.
- 2544 Warren, J.K., 2006, Evaporites: Sediments, Resources, and Hydrocarbons: Berlin, Springer, 1035
2545 p.
- 2546 Weaver, C., and Beck, K., 1977, Miocene of the SE United States – A model for chemical
2547 sedimentation in a peri-marine environment: Sedimentary Geology, v. 17, p. 1-234.
2548

- 2549 Wheeler, J.P., 1954, A geological reconnaissance of the northern Selwyn Mountains region,
2550 Yukon and Northwest Territories: Geological Survey of Canada, Paper 53-7.
2551
- 2552 Wheeler, J.O., and McFeeley, P., 1991, Tectonic assemblage map of the Canadian Cordillera and
2553 adjacent parts of the United States of America: Geological Survey of Canada Map 1712A.
2554
- 2555 Wilson, J.P., Fischer, W.W., Johnston, D.J., Knoll, A.H., Grotzinger, J.P., Walter, M.R.,
2556 McNaughton, N.J., Simon, M., Abelson, J., Schrag, D.P., Summons, R., Allwood, A.,
2557 Andres, M., Gammon, C., Garvin, J., Rashby, S., Schweizer, M., and Watters, W.A., 2010,
2558 Geobiology of the late Paleoproterozoic Duck Creek Formation, Western Australia:
2559 Precambrian Research, v. 179, p. 135-149.
- 2560 Wortmann, U.G., and Paytan, A., 2012, Rapid variability of seawater chemistry over the past 130
2561 million years: Science, v. 337, p. 334-336.
- 2562 Yan, J., Munnecke, A., Steuber, T., Carlson, E.H., and Xiao, Y., 2005, Marine sepiolite in Middle
2563 Permian carbonates of South China: Implications for secular variation of Phanerozoic
2564 seawater chemistry: Journal of Sedimentary Research, v. 75, no. 3, p. 328-338, doi:
2565 10.2110/jsr.2005.026.
- 2566 Yonkee, W.A., Dehler, C.D., Link, P.K., Balgord, E.A., Keeley, J.A., Hayes, D.S., Wells, M.L.,
2567 Fanning, C.M., and Johnston, S.M., 2014, Tectono-stratigraphic framework of
2568 Neoproterozoic to Cambrian strata, west-central U.S.: Protracted rifting, glaciation, and
2569 evolution of the North American Cordilleran margin: Earth Science Reviews, v. 136, no. C,
2570 p. 59-95, doi: 10.1016/j.earscirev.2014.05.004.
- 2571 Young, G.M. 1981, The Amundsen embayment, Northwest Territories: relevance to the upper
2572 Proterozoic evolution of North America, *in* Campbell, F.H.A., editor, Proterozoic basins of
2573 Canada: Geological Survey of Canada Paper 81-10, p. 203-211.
2574
- 2575 Young, G.M., Jefferson, C.W., Delaney, G.D., and Yeo, G.M., 1979. Middle and Upper
2576 Proterozoic evolution of the northern Canadian Cordillera and Shield; Geology, v. 7,
2577 p.125- 128.
2578
2579

2580 FIGURE CAPTIONS

2581 Fig. 1. Simplified location maps of Proterozoic inliers in northwestern Canada adapted
2582 from Young and others (1979), Eisbacher (1981), Wheeler and McFeeley (1991),
2583 Rainbird and others (1996), and Abbott (1997). The blue boxes outline the location of
2584 detailed geological maps presented in figures 3 and 4. This map does not display
2585 Windermere Supergroup strata of the Hyland Group in the Selwyn Basin. NWT–
2586 Northwest Territories; YT–Yukon Territory; SG–Supergroup.

2587

2588 Fig. 2. Schematic lithostratigraphic correlation of Windermere Supergroup strata in
2589 northwestern Canada. The gray box outlines the proposed correlation of the Mount
2590 Harper Group with equivalent strata of the Windermere Supergroup throughout Yukon
2591 and Northwest Territories (NWT). Inset map depicts the rough location of each region
2592 and is adapted from Thomson and others (2014). All italicized names in the stratigraphic
2593 columns lack formalization. ORR—organic-rich-rock; Gp—Group; Fm—Formation; FMG—
2594 Fifteenmile Group; Chan.—Chandindu Formation; Ck.—Creek; Hem. Ck.—Hematite Creek
2595 Group; LD—Little Dal Group; Rav.—Ravensthoat formation; JB—June Beds; GT.—
2596 Gametrail Formation; Mt.—Mount; conglom.—conglomerate; Ft. Co.—Fort Collinson
2597 Formation; Pt.—Point; Hd.—Head.

2598

2599 Fig. 3. Geology of the Hart River Inlier, central Ogilvie Mountains, Yukon. Mapping
2600 based on previous work by Green (1972) and Abbott (1993; 1997) with updates from the
2601 authors over the summers of 2009–2012. Stratigraphic sections plotted on figure 7 and
2602 discussed in the text are depicted as red lines with accompanying section numbers.

2603

2604 Fig. 4. Geology of the Coal Creek Inlier, western Ogilvie Mountains, Yukon, after
2605 Strauss and others (2014b) and references therein. Stratigraphic sections discussed in the
2606 text and plotted on figure 6 are depicted as red lines with accompanying section numbers.

2607

2608 Fig. 5. Photographs of key selected localities from the Coal Creek and Hart River inliers
2609 with solid red lines outlining the trace of measured stratigraphic sections shown in figures

2610 6 and 7. A) Image looking west at the prominent angular unconformity beneath the
2611 Callison Lake Formation in the headwaters of Mark Creek, Hart River inlier (fig. 3). The
2612 distinct erosional unconformity of Rapitan Group strata over the Callison Lake Formation
2613 is also evident. B) Looking west in the NW portion of the Coal Creek inlier at the
2614 profound angular unconformity between the Craggy dolostone (Fifteenmile Group) and
2615 Callison Lake Formation. One can also see the erosional unconformity with the Eagle
2616 Creek Formation (Rapitan Group) cutting down into the Ramp member. C) Impressive
2617 m-scale relief (solid white line) associated with a paleokarst unconformity at the base of
2618 the Callison Lake Formation in the eastern Hart River inlier (section J1223, figs. 3, 7).
2619 Geologist for scale. D) Typical outcrop style of the Callison Lake Formation in the Coal
2620 Creek inlier. Note the discontinuous yellow-orange stromatolitic bioherms in the
2621 Heterolithic member and the prominent black, talc-rich shale interbedded with light gray
2622 dolostone of the Talc member. Tent circled for scale. E) The Harper Fault in the Coal
2623 Creek inlier (fig. 4) as depicted in Mustard and Roots (1997). Largest outlined clast is on
2624 the order of ~50 m tall. F) Looking west in the Coal Creek inlier at the type section of the
2625 Callison Lake Formation (Mount Harper East, J1301) with Mount Harper in the
2626 background. G) Typical outcrop pattern of the Heterolithic member with lenticular bodies
2627 of amalgamated sandstone outlined by dashed white lines. H) Base of the Callison Lake
2628 Fm type section with a decameter-thick stromatolitic “reef” shown above the circled
2629 geologist. The 752.7 ± 5.5 Ma Re-Os age of Rooney and others (2015) is from just above
2630 this stromatolitic unit.
2631

Fig. 6. Detailed measured sections and carbon isotope stratigraphy of the Callison Lake Formation in the Coal Creek inlier. The inset map in the upper left depicts the location of each measured section (GPS coordinates provided in AJS online supplementary data file¹ table DR1). Note the reproducible Islay carbon isotope excursion (ICIE) in the upper Transitional member. The thinning of strata to the NW is a function of the primary accommodation space and secondary erosional truncation during basin-scale uplift in Transitional member and Seela Pass time (see text for an explanation). Strom.—stromatolite; sh/slt—shale/siltstone; sst—sandstone; cglm—conglomerate; dlmud—dolomudstone; dlgrnstn—dolograinstone; dlbdstn—doloboundstone; rxstal—recrystallized.

Fig. 7. Detailed measured sections and carbon isotope stratigraphy of the Callison Lake Formation in the Hart River inlier. The inset map in the upper right depicts the location of each measured section (GPS coordinates in AJS online supplementary data file¹ table DR1). Note the lack of Transitional member strata due to post-Callison Lake erosional truncation beneath the Rapitan Group.

Fig. 8. Selected photographs from Heterolithic member strata in the Hart River and Coal Creek inliers (facies descriptions in table 3). A) Thick-bedded quartz- and chert-pebble conglomerate (Facies, F1) from a basal transgressive lag in the Hart River inlier. Note the abundant jaspillitic chert clasts and yellow-brown oxidized hematitic halos after pyrite. B) Interbedded fine- to medium-grained quartz and chert arenite and gray-green shale and siltstone (F2 and F3). Hammer circled for scale is 27.94 cm tall. C) Mudcracks in green siltstone filled by maroon sandstone (top to left) in the *variegated shale and siltstone*

2655 facies (F3). Canadian coin is 2.39 cm in diameter. D) Dashed white line outlines an
2656 ornately branching stromatolite in the *stromatolitic bioherm and biostrome* facies (F7).
2657 Canadian coin is 2.65 cm in diameter. E) Distinct yellow-orange stromatolitic bioherms
2658 (top outlined with dashed white line) interbedded with gray-green silicified and slightly
2659 cleaved shale and siltstone (F3, 7). Meter stick for scale. F) Domal stromatolites in the
2660 *stromatolitic bioherm and biostrome* facies (F7). Canadian coin is 2.65 cm in diameter.
2661 G) Photomicrograph of disseminated organic matter and dolomite-replaced microbial
2662 sheaths from stromatolitic laminae at Mount Gibben East (section J1019, fig. 6). Scale
2663 bar is 200 μm . H) Dolomite-replaced vase-shaped microfossils (VSMs) in the same
2664 photomicrograph of (G). Scale bar is 150 μm .

2665

2666 Fig. 9. Detailed stratigraphic section of the Talc member, Callison Lake Formation, Coal
2667 Creek inlier (Mine Camp, section J1302). Letters next to the stratigraphic column relate
2668 to the location of accompanying photographs. A) Distinct black, vitreous luster of talc-
2669 rich black shale (Facies, F5; table 1). Canadian coin is 2.65 cm in diameter. B)
2670 Subrounded to rounded talc-rich shale-chip clasts floating in a dolomudstone matrix
2671 (F10). Canadian coin is 2.39 cm in diameter. C) Plan view of large domal stromatolite
2672 (F8) with black, talc-rich shale drapes in individual laminae. Hammer is 27.94 cm tall. D)
2673 Chaotic stromatolitic rudstone with clasts floating in a talc-rich shale matrix. Canadian
2674 coin is 2.39 cm in diameter. E) Ptygmatically-folded microbialite and isopachus laminite.
2675 Hammer is 27.94 cm tall. F) Nodular dolomite-replacement fabric after chickenwire
2676 anhydrite. Canadian coin is 2.65 cm in diameter. G) Microbial dolostone (microbialite)

2677 interbedded with black, talc-rich shale. Canadian coin is 2.39 cm in diameter. Deform.–
2678 deformation; strom.–stromatolite.

2679

2680 Fig. 10. Selected photographs from Ramp member strata in the Hart River and Coal
2681 Creek inliers. A) Trough cross-bedded oolitic dolograinstone (Facies, F11; table 1).
2682 Canadian coin is 2.65 cm in diameter. B) Planar-laminated dolomudstone and dolosiltite
2683 with ferruginous clay partings and abundant black chert nodules (F12) marking a
2684 prominent flooding surface at the base of the Ramp member. Hammer is 31.75 cm tall. C)
2685 Massive and clotty fabric that is reminiscent of microbial fabrics described in Harwood
2686 and Sumner (2011; 2012) from the Beck Spring Dolomite. Canadian coin is 2.65 cm in
2687 diameter. D) Cm-scale oncolites with distinct pendant cements (black arrows with
2688 geopetal features indicating top to the left) floating in a dolosiltite matrix. Canadian coin
2689 is 2.65 cm in diameter. E) High-inheritance domal stromatolites truncated by a distinct
2690 erosional surface (white dashed line) and capped by dolowackestone with abundant
2691 stromatolite clasts (F10). Lens cap is 5.2 cm in diameter. F) Large, m-scale domal
2692 stromatolite buildups (F8). Geologist for scale is 1.75 m tall. G) Partially-silicified
2693 microbialite (F9) characterized by synsedimentary deformation and ptygmatic folding.
2694 Canadian coin is 2.65 cm in diameter. H) Massive silicified dolostone breccia (F14)
2695 characteristic of many paleokarst horizons in the Ramp member. Hammer is 31.75 cm
2696 tall.

2697

2698 Fig. 11. Field photographs of Transitional member strata in the Coal Creek inlier. A)
2699 Digitate stromatolites in the basal transitional member doloboundstone. Canadian coin is

2700 2.65 cm in diameter. B) Irregular, low-inheritance stromatolites with a distinct, healed
2701 synsedimentary normal fault (shown with white dashed line and arrow depicting
2702 hangingwall). Canadian coin is 2.65 cm in diameter. C) Photomicrograph of abundant
2703 vase-shaped microfossils in silicified black shale from Gibben East (section J1204,
2704 Strauss and others, 2014a). Scale bar is 100 μ m. D) Partially silicified massive dolostone
2705 breccia with distinct microbialite clasts from the upper Ramp member. Canadian coin is
2706 2.65 cm in diameter.

2707

2708 Fig. 12. Callison Lake Formation sequence stratigraphic interpretation based on a
2709 reference section from the Coal Creek inlier (Gibben East, composite sections J1018-
2710 J1019; fig. 6). Transgressive-Regressive (T-R) cycles are built upon the sequence
2711 stratigraphic architecture developed by Thomson and others (2015a) in the upper Shaler
2712 Supergroup of Victoria Island and conform to the principles and terminology outlined in
2713 Johnson and Murphy (1984), Johnson and others (1985), Embry and Johannesson (1992),
2714 Embry (2009), and Catuneanu and others (2009; 2011). Lithofacies symbols are the same
2715 as depicted in legends from figures 6 and 7. Ages are from Re-Os geochronology
2716 published in Strauss and others (2014a) and Rooney and others (2015). SU—subaerial
2717 unconformity; MRS—maximum regressive surface; MFS—maximum flooding surface;
2718 Trans.—transgression; Regr.—regression; Fm.—Formation; dolomudstn—dolomudstone;
2719 dologrnstn—dolograinstone; dolobdstn—doloboundstone; rxtal.—recrystallized.

2720

2721 Fig. 13. $\delta^{13}\text{C}_{\text{carb}}-\delta^{18}\text{O}_{\text{carb}}$ cross-plots for individual members of the Callison Lake
2722 Formation.

2723

2724 Fig. 14. Cartoon depositional reconstruction and schematic tectono-stratigraphic
2725 evolutionary model for the Callison Lake Formation. Note that only a small portion of
2726 these hypothetical basins are exposed in the Coal Creek and Hart River inliers. A)
2727 Initiation of Callison Lake sedimentation is recorded in syn-rift mixed siliciclastic,
2728 evaporite, and carbonate deposits of the Heterolithic and Talc members, which were
2729 deposited in marginal marine embayments associated with an ancestral basin-bounding
2730 extensional fault system. B) Ramp member deposition records a shift to pure carbonate
2731 sedimentation and the development of a hangingwall dip-slope ramp. Detailed inset
2732 depicts a hypothetical reconstruction of the inner ramp setting with characteristic
2733 depositional environments for specific Ramp member facies (table 3). Modified from
2734 Harwood and Sumner (2011). C) Transitional member sedimentation records a change in
2735 Mount Harper Group basin dynamics and the onset of renewed extensional tectonism.
2736 This is evidenced by coeval subaerial exposure and marine sedimentation associated with
2737 rotation of the Callison Lake hangingwall depocenter, the northward propagation of
2738 basin-bounding structures accompanied by segmentation of the original Callison Lake
2739 basin, and the eventual progradation of fault-related siliciclastic deposits of the Seela
2740 Pass Formation. This event may also record the onset of strike-slip tectonism throughout
2741 the Ogilvie and Mackenzie Mountains as depicted by the activation of the Dawson Fault.

2742

2743 Fig. 15. Preliminary lithostratigraphic, chemostratigraphic, and sequence stratigraphic
2744 correlations among basal Windermere Supergroup strata in northwestern Canada adapted
2745 from Strauss and others (2014a). Note that the Islay carbon isotope excursion (ICIE) is

2746 consistently present in the third Transgressive-Regressive (T-R) cycle (T-R8). Schematic
2747 stratigraphy, chemostratigraphy, and geochronology of each column are from the
2748 following sources: Ogilvie Mountains (Macdonald and others, 2010; Strauss and others,
2749 2014a; Rooney and others, 2015); Mackenzie Mountains (Jefferson, ms, 1983; Jefferson
2750 and Parrish, 1989; Rooney and others, 2014; Milton and others, 2015); Victoria Island
2751 (Rainbird, 1993; Long and others, 2008; Jones and others, 2010; Macdonald and others,
2752 2010; van Acken and others, 2013; Prince, ms, 2014; Thomson and others, 2015b).
2753 RAP.–Rapitan Group; Mtns–Mountains; MHV–Mount Harper Volcanics; 15 Mi.–
2754 Fifteenmile Group; NWT–Northwest Territories; Fm.–Formation; LDB–Little Dal basalt;
2755 Natkus.–Natkusiak Formation; W–Wynniatt Formation.

GEOLOGY OF THE HART RIVER INLIER

STRUCTURE - CONTACTS

— defined
 --- approximate
 - - - inferred

STRUCTURE - FAULTS/FOLDS

— normal, defined
 - - - normal, approximate
 - - - normal, inferred
 - - - thrust, defined
 - - - thrust, approximate
 - - - thrust, inferred
 - - - unknown, defined
 - - - unknown, approximate
 - - - unknown, inferred
 - - - syncline
 - - - anticline

OTHER

— measured sections
 - - - edge of detailed mapping
 ○ waterbody

STRATIGRAPHY

TRIASSIC

Tg: GALENA SUITE

CARBONIFEROUS

Mkhu: KENO HILL QUARTZITE

DEVONIAN

DE: EARN GROUP

Do: OGILVIE Fm

ORDOVICIAN - DEVONIAN

ODr: ROAD RIVER Gp: shale
 ODrd: ROAD RIVER - Duo Lake Fm

ODs: SAPPER Fm

CAMBRIAN - ORDOVICIAN

COdv: DEMPSTER VOLCANICS

COvu: Paleozoic volcanics undivided

COdu: dykes undivided

COgl: GULL LAKE Fm

COrr: RABBITKETTLE Fm

HYLAND GP (EDIACARAN - L. CAMBRIAN)

nPChn: NARCHILLA Fm

nPha: ALGAE Fm

nPhy: YUSEZYU Fm

nPhu: HYLAND Gp undivided

NEOPROTEROZOIC

HAY CREEK/UPPER GP (CRYOGENIAN - EDIACARAN)

nPHCu: HAY CREEK/UPPER Gps undivided

nPHCv: Ediacaran(?) volcanics undivided

RAPITAN GP (CRYOGENIAN)

nPRu: RAPITAN Gp undivided

MOUNT HARPER GP (TONIAN)

nPHu: MOUNT HARPER Gp undivided

nPHd: CALLISON LAKE Fm

FIFTEENMILE GP (TONIAN)

nPFu: FIFTEENMILE Gp undivided

nPFppd3: FIFTEENMILE Gp - PPD3

nPFppd2: FIFTEENMILE Gp - PPD2

nPFppd1: FIFTEENMILE Gp - PPD1

MESOPROTEROZOIC

PINGUICULA GP (MESOPROTEROZOIC - NEOPROTEROZOIC?)

PPu: PINGUICULA Gp undivided

PPa: PINGUICULA Gp - Unit A

PPb: Pinguicula Gp - Unit B

PPc: Pinguicula Gp - Unit C

WERNECKE SUPERGROUP (MESOPROTEROZOIC)

PGL: GILLESPIE LAKE Gp

PGL1: GILLESPIE LAKE Gp: shale

PHRs: HART RIVER sills

PHRv: HART RIVER volcanics

PQ: QUARTET

CALLISON LAKE FORMATION (Hart River Inlier) Heterolithic, Talc, Platformal, and Transitional members

Rapitan Group Undivided
(Map Units Hv, Hs, Hdi of Abbott, 1997)

**MINE CAMP
SECTION
J1302**
Talc member
of the
Callison Lake
Formation
(detailed section)

Heterolithic (n = 105)

Platformal (n = 648)

Talc (n = 339)

Transitional (n = 130)

A. Heterolithic and Talc member time (ca. 780(?)–750 Ma)

B. Platformal member time (ca. 750–740 Ma)

C. Transitional member and Seela Pass time (ca. 740–720 Ma)

TABLE 1. SUMMARY OF CALLISON LAKE FORMATION LITHOFACIES (PAGE 1)

Lithofacies	Composition	Bedding Style/Structures	Depositional Environment	Distribution
Siliciclastic Facies:				
<u>F1: Pebble to granule conglomerate</u>	Clast- to matrix-supported conglomerate; light grey to brown; dominated by quartz and chert with occasional lithics; well-rounded to subangular; moderate sorting; hematite and silica cement	Thin- to thick-bedded; occasionally forming distinct transgressive lags; faint ripple- to dune-scale trough and tabular cross-bedding up to 20 cm thick; erosional base, lenticular geometry, and fining-up packages common; associated with F2 and F3	Braided fluvial channels. Tidal/Estuarine? Amalgamated channel deposits with abrupt lateral facies change; difficult to discern distinct point-bar or lateral accretion geometries; locally reworking underlying strata; transgressive lag	Restricted to HM
<u>F2: Sandstone</u>	Quartz and chert arenite and wacke; very fine- to coarse-grained with occasional granules; moderate sorting; well-rounded to subangular; abundant hematite staining; tan yellow to brown; silica cement	Thin- to medium-bedded; commonly with unidirectional and symmetrical ripple cross-bedding and parallel lamination; occasionally amalgamated; mud chip intraclasts common; interbedded with F3 and forming fining-up packages with F1	Coastal plain/Estuarine. Maybe local shoreface? Evidence for floodplain or tidal flat deposition with clear tidal influence in the form of distinct tidal ravinement surfaces and fining-up packages	Restricted to HM
<u>F3: Variegated siltstone and shale</u>	Variegated shale (red, yellow, green, and purple) interbedded with siltstone; locally contains diagenetic dolostone lenses and iron formation; heavily silicified in certain horizons; occasional very-fine sand sized particles	Locally abundant mudcracks and scours; mostly planar-laminated and associated with F1 and F2; occasional coarsening-upward packages with ripple cross lamination; flaser-bedding and ball and pillows locally	Floodplain or peritidal mud flat/lagoon. Generally associated with transition into marginal marine inner ramp to lagoonal deposits; interbedded sandstone units could represent crevasse splays in a fluvial setting	Restricted to HM
<u>F4: Black shale</u>	Black to dark grey shale; TOC to 3.5 wt. %; occasionally silicified with abundant chert nodules; locally contains vase-shaped microfossils; minor silt and very-fine sand; Fe-oxides, pyrite, sphalerite, ankerite(?) locally	Planar-laminated and fissile; occasionally silicified and more resistant; locally interbedded with various lithofacies and no evidence for wave or storm activity; poor sorting with coarser grained horizons	Outer-inner ramp subtidal or lagoonal. Suspension depositor in a low-energy setting; elevated TOC tied to episodic restriction?	HM, TRM
Evaporite/Talc Facies:				
<u>F5: Talc-rich shale</u>	Black talc-rich shale; commonly silicified with abundant chert nodules; organic-rich with TOC up to 4 wt%	Planar-laminated to nodular bedded; abundant diagenetic chert; occasional soft-sediment deformation; intimate association with F6	Restricted lagoonal to sahbka. Playa lake? Suspension deposition in a low-energy setting; evidence for seismic or storm-generated disruption	TM, TRM?
<u>F6: Interbedded talc and dolostone</u>	Black talc-rich shale interbedded with microbial and sucrosic light to dark dolostone; commonly silicified with chert nodules	Nodular- to planar-bedded with microbial lamination; teepees structures, mudcracks, and evaporite pseudomorphs after anhydrite and gypsum common; horizons of talc-shale chip intraclast conglomerate common; local overturned stromatolites and seismites	Sahbka to peritidal mud flat. Evidence for intertidal to peritidal deposition; episodic restriction with sulfate deposition; wave-, storm-, and seismically-generated structures common; complex paragenetic history with multiple Mg-silicate and evaporite transformations	Restricted to TM
Carbonate Facies:				
<u>F7: Stromatolitic bioherm/biostrome</u>	Orange-yellow to dark grey (fresh) stromatolitic doloboundstone; minor terrigenous silt and frosted quartz grains; occasional vase-shaped microfossils in disseminated organic matter; microbial sheaths after cyanobacteria	Meter- to decimeter-thick stromatolitic biohermal/biostromal buildups interbedded with F3,4; stromatolites range from domal to columnar with both low- and high-inheritance forms; micritic, oolitic, and intraclastic fill; isolated to laterally linked forms	Middle- to inner-ramp subtidal or lagoonal. Isolated to linked stromatolitic patch reefs or buildups associated with a subtidal depositional setting; possibly lagoonal with no evidence for exposure or wave activity	HM, TRM

HM – Heterolithic member, TM – Talc member, PM – Platform member, TRM – Transitional member

TABLE 1. SUMMARY OF CALLISON LAKE FORMATION LITHOFACIES (PAGE 2)

Lithofacies	Composition	Bedding Style/Structures	Depositional Environment	Distribution
<u>F8: Stromatolitic doloboundstone</u>	Light to dark grey stromatolitic doloboundstone; occasional black chert and ferruginous clay-rich laminae; close association with stromatolite-rich intraclast wackestone and grainstone	Centimeter- to meter-thick stromatolitic structures; forms range from laterally-linked, low relief, and domal to high-inheritance columnar structures; intercolumnal fill micritic, oolitic, pisolitic, and intraclasts; scours and corrosive structures common	Middle-inner ramp subtidal. Heterogeneous assemblage of stromatolitic morphologies; common wave-generated scours and association with F9 and F10; meter-scale high relief domes suggest occasional deeper water setting	All members.
<u>F9: Microbialite</u>	Light to dark grey microbial doloboundstone; occasional black chert nodules, terrigenous silt, fine- to medium-sized frosted quartz grains; distinguished from F8 by flat, crinkly lamination	Centimeter- to meter-thick microbialite; mm-scale undulatory and wavy lamination; locally containing fenestrae (birds' eye), teepee structures, and mudcracks; occasional seismically-disrupted and folded laminae; discrete intervals of intraclast conglomerate	Inner ramp intertidal to peritidal. Intertidal deposition evidenced by intimate association with F8 and wave- or storm-generated intraclasts; exposure surfaces common; discrete parasequences with F6,8, and 10	All members.
<u>F10: Intraclast grainstone/wackestone</u>	Light to dark grey dolomitic intraclast grainstone and wackestone; almost exclusively composed of clasts from F8 and F9; minor rudstone	Thin- to thick-bedded, tabular to subrounded, sand- to cobble-sized intraclastic grainstone and wackestone matrix-supported with scoured and erosive bases, occasional crude lamination, and faint grading; commonly randomly oriented	Inner ramp subtidal. Subtidal deposits associated with high-energy storm- or wave-generated events that scour and rework material from F9 and F10; no evidence for subaerial exposure	PM, TRM
<u>F11: Dolograinstone</u>	Light to dark grey dolograinstone composed of ooids, peloids, pisoids, and/or oncoids; minor terrigenous silt component	Thin- to thick-bedded and massive to finely laminated; occasional trough cross-bedding and planar laminated but generally quite massive and crudely stratified; intimately associated with F8, F9, and F10; occasional low-angle cross-bedding	Inner ramp subtidal to intertidal bar or shoal complex. High-energy subtidal deposition associated with tidal sand bars and/or migrating barrier complex; 3D morphology not worked out but commonly interbedded with diverse lithofacies	Restricted to PM
<u>F12: Thin-bedded dolomicrite/siltite</u>	Dark grey to black dolomicrite and dolosiltite; commonly composed of microcrystalline and neomorphosed dolospar and dolomicrite; occasional hints of peloidal precursor allochems; ferruginous clay drapes and black chert common	Thin-bedded and laminated with mm-scale parallel lamination; occasional erosional base; black chert commonly late diagenetic and fabric destructive; no evidence for grading or turbidite deposition; commonly resemble Phanerozoic ribbon-bedded limestone	Mid-outer ramp subtidal. Generally constrained to transgressive horizons associated with relatively low-energy deposition; ferruginous drapes could indicate hardground conditions or suspension rainout during highstand; occasionally deformed during seismic activity	PM, TRM
Diagenetic Facies: <u>F13: Recrystallized dolostone</u>	Light grey to white sucrosic dolostone with fabric-destructive diagenetic recrystallization; occasional oolitic ghosts	Medium- to thick-bedded and massive; crude stratification and commonly impossible to tell precursor lithology; abundant secondary isopachous, drusy, and botryoidal cements; veins common; locally contains clotty secondary fabric that resembles "thrombolitic" texture	Unclear but commonly associated with subaerial exposure surfaces and coarser-grained lithofacies.	All members
<u>F14: Breccia</u>	Grey to white dolostone breccia; commonly silicified with angular clasts of dolostone; matrix composed of both terrigenous silt, secondary dolomite spar, and dolosiltite; clasts range from sand- to boulder-sized; local well-rounded quartz sand within cavity fill	Massive and generally thick-bedded; crude stratification with clear horizons indicating subaerial exposure and karst development; clasts composed exclusively of underlying strata in puzzle-fitting fabric; local grykes and irregular cavities filled with terrigenous silt and sand	Inner ramp intertidal to peritidal. Clearly associated with subaerial exposure and karst development in dolostone lithologies; possibly associated with uplift and fault breccia in certain localities	PM, TRM

HM – Heterolithic member, TM – Talc member, PM – Platform member, TRM – Transitional member

STRAUSS AND OTHERS (XXX) SUPPLEMENTARY DATA FILE XXX

Table DR1. Formalization of the Seela Pass Formation.

Name	Seela Pass Formation
Name Derivation	Type area located west of Seela Pass, western Ogilvie Mountains, Yukon Territory, Canada; Dawson Quadrangle (NTS 116BC)
Category and Rank	lithostratigraphic Formation
Type Area	Situated broadly between Eagle Creek and Chandindu River, western Ogilvie Mountains, Yukon Territory, Canada
Unit Type Section	Composite stratigraphic sections 10 and 12 of Mustard (1991) Located on prominent N-S trending ridgeline ~8.7 km ENE of Mount Harper (N64°39.539' W139°44.405') Lower boundary: gradational transition from Transitional member of Callison Lake Formation Upper boundary: covered, sharp transition into Mount Harper Volcanics
Unit Description	Mustard (1990; 1991) documented five main facies in the Seela Pass Formation including <i>fault-adjacent breccias</i> , <i>coarse conglomerate</i> , <i>conglomerate-sandstone</i> , and <i>mudstone-sandstone</i> . The type section consists of ~1100 m of mostly thick-bedded, sheet-like, poorly sorted and disorganized massive conglomerate. The conglomerate consists of clast-supported, pebble- to boulder-sized clasts of dolostone and sandstone (with minor chert) in a dolowacke matrix with minor matrix-supported lenses and sandstone. Grading is poorly developed and clasts are typically subangular to subrounded. Near the Harper Fault, there are distinct lenses of clast-supported dolostone megaboulders > 20 m across. Further to the east, the Seela Pass Formation generally becomes finer-grained and displays a distinct gradational transition in bedding thickness and grain-size into the other characteristic facies of the unit. The other reference section (section 20 of Mustard, 1991) is dominated by maroon dolomitic mudstone and siltstone with mudcracks that coarsen-upwards into thicker deposits of sandstone and minor conglomerate. Overall, Mustard (1991) interpreted the Seela Pass Formation as a distinct progradational alluvial fan and fan delta succession.
Unit Reference Sections	Stratigraphic section 20 of Mustard (1991) Located in a small N-facing gully ~18 km from Seela Pass (N64°40.946 W139°13.982)
Dimensions	~1100 m thick in composite section at type section (Mustard, 1991) ~605 m at reference section (Mustard, 1991)
Geologic Age	Neoproterozoic (<739.9±6.1 Ma, >717.43±0.14 Ma, Macdonald et al., 2010; Strauss et al., 2014a)
Regional Correlations	unnamed equivalent in Hart River inlier; Coates Lake Group, Mackenzie Mountains; Kuujjua Formation, Shaler Supergroup

Table DR2. Formalization of the Callison Lake Formation.

Name	Callison Lake Formation
Name Derivation	Type area located in the Coal Creek inlier, Ogilvie Mountains, Yukon Territory, Canada; Dawson Quadrangle (NTS 116BC)
Category and Rank	lithostratigraphic Formation
Type Area	Situated broadly between Eagle Creek and Chandindu River, western Ogilvie Mountains, Yukon Territory, Canada
Unit Type Section	Mount Harper East, section J1301 (Figure 6; <i>this paper</i>)
Unit Description	<p>Located on prominent N-S trending ridgeline ~10 km ENE of Mount Harper</p> <p>Lower boundary: sharp contact on silicified karst of Craggy dolostone, Fifteenmile Group (N64.6644833°, W-139.73135°)</p> <p>Upper boundary: gradational transition into Seela Pass Formation (N64.6599667°, W-139.73685°)</p> <p>Divided into four informal members: (1) Heterolithic member: heterogeneous siliciclastic-dominated package of interbedded conglomerate, sandstone, siltstone, and shale with minor stromatolitic and microbial dolostone. Conglomerate is both clast- and matrix-supported and dominated by pebble- to cobble-size clasts of quartz, chert, and occasional lithics. Sandstone units are characterized by subangular to subrounded quartz and chert arenite with minor wacke. Coarse lithologies tend to be thick- to medium-bedded and host trough- and tabular-crossbedding, soft sediment deformation, and erosional scours. Shale and siltstone units are generally variegated, locally contain mudcracks and synaeresis cracks, and locally become organic-rich. Yellow-brown stromatolitic dolostone units generally form prominent biostromes or bioherms and are laterally discontinuous. Lower contact is sharp and erosive on paleokarst intervals of upper Craggy dolostone and upper contact is sharp with overlying strata of Talc member: (2) Talc member: interbedded black, talc-rich shale and grey-white dolostone. Talc-rich shale displays a vitreous luster, is locally finely laminated and pure, but also drapes microbial dolostone units and is commonly interbedded with nodular dolostone. Dolostone units are characterized by stromatolitic and microbial doloboundstone with intervals of wackestone, grainstone, and rudstone. Common sedimentary structures include evaporitive pseudomorphs, mudcracks, teepees, crudley laminated intraclast rip-ups, and occasional erosional scours. Upper contact is sharp with the overlying Platform member dolostone. (3) Platform member: dominated by light- to dark-grey, medium-thick-bedded dolograins, doloboundstone, dolowackestone, and dolorudstone. Allochems characterized by pisoids, ooids, peloids, oncoids, and distinct stromatolite clasts ("flakestone") and common sedimentary structures include dune-scale crossbedding, erosional scours within doloboundstone units, morphologically-diverse stromatolites and microbialite, occasional evaporite pseudomorphs, teepees, and massively recrystallized intervals. Upper contact either gradational into Transitional member deposits or marked by a profound subaerial exposure surface and paleokarst interval. (4) Transitional member: heterogeneous package of interbedded microbial and stromatolitic dolostone and black shale/siltstone. Dolostone units marked by morphologically-diverse stromatolites, healed synsedimentary faults, seismites subaerial exposure surfaces, teepees, and abundant detrital material. Siliciclastic intervals commonly fine-grained, organic-rich, silicified, and occasionally poorly sorted. Upper contact gradational into siliciclastic deposits of the overlying Seela Pass Formation.</p>
Unit Reference Sections	<p>1. Mount Gibben, composite section J1018-1019 (Figure 6, <i>this paper</i>)</p> <p>Located on a prominent N-S trending ridgeline with upper part of section exposed to SW (N64.6886°, W-139.3528167°)</p> <p>2. Sheep Camp, section J907, Hart River inlier (Figure 7, <i>this paper</i>)</p> <p>Located in a steep, N-facing gully and SW-trending ridgeline (64.572229°, -136.836461°)</p>
Dimensions	<p>409.6 m thick at type section (Figure 6, <i>this paper</i>)</p> <p>516.4 m thick at reference section (Figure 7, <i>this paper</i>)</p>
Geologic Age	<p>Neoproterozoic (<780 Ma, >717.43±0.14 Ma, Macdonald et al., 2010; Strauss et al., 2014a; <i>this paper</i>)</p> <p>Two internal Re-Os depositional ages: 752.7±5.1 Ma (Heterolithic member), 739.9±6.1 Ma (Transitional member) (Strauss et al., 2014a; Rooney et al., in review)</p>
Regional Correlations	Thundercloud, Redstone River, and Coppercap formations of Coates Lake Group, Mackenzie Mountains; Kilian and Kuujjua formations, Shaler Supergroup

Table DR3. Location of measured stratigraphic sections.

Section Name	Camp Name	Inlier	Latitude	Longitude
F927 - base	Talc Falls	Coal Creek	64.787756	-139.825689
F927 - top	Talc Falls	Coal Creek	64.787531	-139.828554
F928 - base	Talc Falls	Coal Creek	64.781026	-139.822763
F928 - top	Talc Falls	Coal Creek	64.777922	-139.836666
J907 - base	Sheep Camp	Hart River	64.572229	-136.836461
J907 - top	Sheep Camp	Hart River	64.573667	-136.849934
J908 - base	Sheep Camp	Hart River	64.569988	-136.802585
J908 - top	Sheep Camp	Hart River	64.568602	-136.803404
J1018 - base	Mount Gibben	Coal Creek	64.6631833	-139.3974333
J1018 - top	Mount Gibben	Coal Creek	64.6627833	-139.3969833
J1019 - base	Mount Gibben	Coal Creek	64.6886	-139.35081
J1019 - top	Mount Gibben	Coal Creek	64.6725	-139.3692667
J1122 - base	Mark Creek	Hart River	64.623077	-136.930764
J1122 - top	Mark Creek	Hart River	64.620149	-136.927846
J1201 - base	Gibben East	Coal Creek	64.70755	-139.1039333
J1201 - top	Gibben East	Coal Creek	64.7064667	-139.1080333
J1202 - base	Gibben East	Coal Creek	64.6986667	-139.0937167
J1202 - top	Gibben East	Coal Creek	64.69785	-139.09945
J1203 - base	Gibben East	Coal Creek	64.69685	-139.0962333
J1203 - top	Gibben East	Coal Creek	64.69655	-139.0963833
J1204 - base	Gibben East	Coal Creek	64.68155	-139.2321
J1206 - base	Reefer Camp	Coal Creek	64.7568833	-139.7067667
J1206 - top	Reefer Camp	Coal Creek	64.758063	-139.752754
J1207 - base	Reefer Camp	Coal Creek	64.7617667	-139.7470167
J1207 - top	Reefer Camp	Coal Creek	64.7587	-139.7526
J1210 - base	Mount Harper	Coal Creek	64.6883333	-139.85805
J1210 - top	Mount Harper	Coal Creek	64.6829333	-139.8633667
J1211 - base	Mount Harper	Coal Creek	64.6920333	-139.8670333
J1211 - top	Mount Harper	Coal Creek	64.6905833	-139.867
J1222 - base	Penetration Lake	Hart River	64.54539	-136.549296
J1222 - top	Penetration Lake	Hart River	64.542626	-136.541623
J1223 - base	Penetration Lake	Hart River	64.544011	-136.487438
J1223 - top	Penetration Lake	Hart River	64.543031	-136.484123
J1301 - base	Mount Harper East	Coal Creek	64.6644833	-139.731729
J1301 - top	Mount Harper East	Coal Creek	64.6599667	-139.73685
J1302 - base	Mine Camp	Coal Creek	64.8242	-140.0427333
J1302 - top	Mine Camp	Coal Creek	64.8276833	-140.0504333
J1303 - base	Mine Camp	Coal Creek	64.8148333	-140.0499
J1303 - top	Mine Camp	Coal Creek	64.8126667	-140.0478333

Table DR4. Carbonate carbon and oxygen isotope data.

Section	Stratigraphic Height	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	Section	Stratigraphic Height	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$
J1302	1.2	4.51	0.52	J1018	0.1	4.21	1.74
J1302	2.3	4.04	-1.39	J1018	0.5	3.78	0.20
J1302	6.8	4.88	-1.07	J1018	1	3.35	-1.29
J1302	8.6	3.21	-2.57	J1018	1.5	3.59	-0.95
J1302	10.7	-0.68	-1.84	J1018	2	3.51	-0.87
J1302	14.6	3.76	-1.43	J1018	2.5	3.32	-1.48
J1302	16.5	3.54	-1.72	J1018	3	3.33	-1.51
J1302	18.3	2.96	-1.45	J1018	3.5	3.32	-0.89
J1302	20.7	2.37	3.37	J1018	4.5	3.93	0.97
J1302	22.8	3.99	0.40	J1018	5	4.19	1.32
J1302	25.5	1.10	2.89	J1018	5.5	3.81	-0.22
J1302	28.1	1.57	-0.75	J1018	6	3.62	0.24
J1302	30.8	-0.10	2.43	J1018	6.5	3.16	-0.63
J1302	32.5	-1.50	3.62	J1018	7	2.63	0.30
J1302	33.5	-2.92	3.62	J1018	7.5	2.14	-0.84
J1302	35.3	1.10	2.06	J1018	8	2.28	-0.87
J1302	36.9	3.37	-1.78	J1018	9.2	1.17	-0.20
J1302	37.8	3.11	-0.20	J1018	9.5	0.82	-1.05
J1302	41.6	-2.01	-3.47	J1018	10	0.68	-2.48
J1302	43.8	3.17	-0.71	J1018	10.5	0.73	0.33
J1302	46.4	3.38	-0.88	J1018	11	0.99	0.15
J1302	47.3	1.92	-2.34	J1018	11.5	1.29	-0.13
J1302	51.3	0.57	-1.76	J1018	12	1.26	0.74
J1302	54.6	-0.80	0.32	J1018	12.5	1.86	1.56
J1302	57.8	-1.33	3.76	J1018	13	1.49	-0.75
J1302	59.6	1.77	-0.37	J1018	14	0.61	0.88
J1302	62.6	2.65	1.38	J1018	14.5	-0.33	0.16
J1302	64.6	1.78	-0.51	J1018	15	1.92	1.32
J1302	66.6	2.48	-0.03	J1018	16	1.27	0.04
J1302	68.6	2.28	-2.60	J1018	16.5	0.48	0.70
J1302	72.2	0.75	3.21	J1018	17	0.45	1.25
J1302	72.8	0.43	0.56	J1018	17.5	-0.06	1.43
J1302	75.6	-0.62	1.86	J1018	18	-0.56	0.58
J1302	78.5	-1.35	-0.34	J1018	18.5	-0.40	1.45
J1302	79.6	-0.61	-1.12	J1018	19	-0.73	-0.11
J1302	81.7	-0.66	-0.56	J1018	19.5	-0.19	0.66
J1302	83.8	-0.72	-3.31	J1018	20	-2.01	-0.81
J1302	84.4	-0.18	-2.36	J1018	28	-1.74	1.57
J1302	109	-0.66	-1.22	J1018	29	-0.47	-1.04
J1302	111.1	-1.81	0.32	J1018	29.5	-1.23	-3.77
J1302	113.4	0.02	-1.95	J1018	30	-3.27	-0.62
J1302	125.9	-1.23	-2.51	J1018	30.5	-2.00	0.64
J1302	129.8	0.58	-2.33	J1018	31	-0.94	0.15
J1302	134.8	2.63	-0.64	J1018	31.5	-0.35	-2.17
J1302	139.7	1.67	-1.07	J1018	32	-0.61	-2.66
J1302	144.8	2.69	1.19	J1018	37.1	-2.83	-1.46
				J1018	37.5	-2.43	-1.75
J1201	74.4	1.61	-6.80	J1018	38.5	-1.56	-0.93
J1201	74.8	-0.39	-6.70	J1018	39	-2.31	-1.44
J1201	75.6	-0.08	-7.75	J1018	39.5	-1.20	-0.31
J1201	76.2	0.28	-6.99	J1018	40	-1.03	-0.66
J1201	76.5	-0.13	-5.89	J1018	40.5	-1.01	-0.62
J1201	77.5	1.42	-5.73	J1018	41	-0.53	-0.76
J1201	77.6	0.64	-6.10	J1018	41.5	0.18	-0.90
J1201	78	0.38	-6.02	J1018	42	-0.36	-1.07

J1201	78.3	0.38	-6.21	J1018	42.6	0.77	-1.01
J1201	86.5	0.36	-6.47	J1018	43	0.37	-1.03
J1201	87.1	0.73	-7.31				
J1201	87.5	1.55	-7.19	J1210	0.6	1.81	-3.75
J1201	88	1.78	-7.33	J1210	2.3	2.16	-2.86
J1201	88.5	0.53	-6.92	J1210	3.2	2.36	-3.04
J1201	89	1.26	-6.93	J1210	4.1	1.50	-2.34
J1201	89.5	1.32	-6.89	J1210	6.1	2.57	-1.21
J1201	90	0.76	-7.15	J1210	8.2	2.74	-2.45
J1201	90.5	0.63	-6.88	J1210	9.1	3.42	-1.68
J1201	90.9	0.75	-7.15	J1210	10.1	2.87	-2.34
J1201	91.2	0.82	-6.93	J1210	11.1	3.39	-4.96
J1201	92.5	0.42	-7.45	J1210	12.6	4.20	-3.73
J1201	93	-0.47	-7.30	J1210	13.9	5.07	-2.53
J1201	93.8	-0.35	-6.98	J1210	15	5.33	-1.78
J1201	94.5	0.08	-6.64	J1210	19.2	5.84	-3.04
J1201	96	0.68	-6.84	J1210	22.6	6.09	1.43
J1201	96.3	0.39	-6.63	J1210	24.4	5.50	-3.10
J1201	96.7	1.12	-6.77	J1210	26.3	5.97	-0.01
J1201	107.5	-1.58	-4.33	J1210	28.4	5.33	-3.08
J1201	108.1	1.54	-3.43	J1210	38.8	3.10	-2.08
J1201	108.5	0.80	-2.94	J1210	41	3.99	-4.17
J1201	109.1	1.14	-2.83	J1210	43	4.55	-0.56
J1201	109.4	-0.20	-3.86	J1210	61.6	4.97	0.96
				J1210	62.1	2.33	1.88
J1202	0.3	4.11	-1.16	J1210	63.2	5.08	-2.60
J1202	1.8	3.86	0.47	J1210	85.4	4.44	-2.94
J1202	3.9	3.61	-1.20	J1210	87.1	4.59	1.36
J1202	11.4	0.68	0.83	J1210	90.2	4.11	-5.07
J1202	27.6	1.53	-5.32	J1210	93.4	4.01	-4.42
J1202	36.8	2.20	-0.36	J1210	95.1	2.48	-0.70
J1202	48.7	2.10	-1.96	J1210	96.5	2.52	0.92
J1202	58.1	1.54	-1.57	J1210	97.6	5.00	2.07
J1202	58.3	2.17	0.45	J1210	99.2	3.82	0.63
J1202	61.8	2.90	0.23	J1210	101.4	4.15	0.99
J1202	74.3	1.54	-6.17	J1210	103.1	4.53	1.34
J1202	81.6	2.02	-5.87	J1210	105	3.73	-4.64
J1202	86.7	2.06	-4.54	J1210	106.2	3.93	-5.92
J1202	109.2	1.92	-9.05	J1210	108.9	2.18	-0.99
J1202	119.7	1.41	-6.13	J1210	109.8	3.54	-4.67
J1202	130.2	2.19	-5.84	J1210	107.5	2.70	-2.51
J1202	155.4	1.32	-6.07	J1210	110.7	3.68	-1.17
J1202	166.2	1.26	-5.72	J1210	113.3	4.26	0.58
J1202	176.7	1.29	-5.32	J1210	114.9	4.61	0.04
J1202	187.2	-0.48	-2.25	J1210	117.1	5.03	1.84
J1202	197.7	0.13	-0.72	J1210	119.2	4.61	2.26
J1202	208.2	0.91	-3.56	J1210	123.1	4.46	2.79
J1202	218.5	3.63	-0.43	J1210	127.2	4.88	1.60
				J1210	129.2	5.39	2.25
J1203	0.3	0.56	-6.23	J1210	133.4	5.83	2.24
J1203	0.9	0.64	-7.05	J1210	137.8	5.63	1.89
J1203	1.6	0.44	-7.59	J1210	140.3	5.87	2.49
J1203	2.5	1.21	-6.06	J1210	142.4	6.05	1.96
J1203	3.3	1.05	-6.36	J1210	147.1	5.75	1.31
J1203	4.1	1.25	-5.19	J1210	149.5	5.70	1.50
J1203	4.15	1.54	-5.10	J1210	152.5	5.96	1.46
J1203	5	1.34	-5.10	J1210	155.1	5.81	1.05
J1203	6.3	1.39	-6.43	J1210	157.1	5.56	-1.19

J1203	7	1.93	-3.93	J1210	159.4	4.96	0.82
J1203	8.8	2.05	-4.96	J1210	163	5.26	0.92
J1203	9.7	2.71	-5.49	J1210	165.4	5.50	1.92
J1203	10.1	3.33	-1.37	J1210	168.1	5.87	4.19
J1203	12.1	3.64	-0.37	J1210	172.3	5.48	0.47
J1203	13.2	2.10	-4.69	J1210	174.1	5.02	-2.13
J1203	14.2	1.29	-5.02	J1210	176	4.77	1.05
J1203	15.1	0.99	-1.54	J1210	177.9	4.62	0.62
J1203	16	0.90	-0.70	J1210	180.2	4.30	-0.76
J1203	17	-0.84	-2.22	J1210	182.2	4.72	0.39
J1203	17.8	-2.49	0.68	J1210	183.7	4.84	2.13
J1203	18.8	-1.63	0.24	J1210	186.1	5.06	0.71
J1203	21.1	-3.28	0.31	J1210	187.9	4.91	0.62
J1203	25.2	-4.18	-1.57	J1210	192.3	4.73	-1.77
J1203	25.4	-3.74	0.07	J1210	194.2	4.89	-0.29
J1203	35.5	-0.73	-0.63	J1210	196.2	4.22	0.59
J1203	36.1	-1.41	-2.21	J1210	197.8	2.02	0.64
J1203	36.5	-2.10	-1.49	J1210	200.1	4.68	-2.97
				J1210	201.9	4.66	-3.11
J1204	0.3	4.00	-2.84	J1210	204.1	4.16	-5.72
J1204	1	3.91	-3.11	J1210	207.8	4.42	-3.08
J1204	1.3	3.61	-5.12	J1210	206.1	4.74	-3.23
J1204	2	3.77	-4.16	J1210	206.2	5.03	-0.78
J1204	2.5	3.61	-4.36	J1210	208.2	4.29	-1.04
J1204	3	3.78	-2.45	J1210	210.7	4.58	0.63
J1204	3.4	3.15	-2.96	J1210	212.1	4.82	-1.28
J1204	4.1	2.58	0.35	J1210	214.1	4.53	1.92
J1204	4.4	2.26	-0.19	J1210	216.2	5.15	0.48
J1204	5	2.01	-0.93	J1210	218	4.53	-1.50
J1204	5.5	1.67	0.42	J1210	220.1	4.66	-0.17
J1204	6.1	1.63	0.11	J1210	222.4	4.92	-0.44
J1204	6.5	1.29	-0.19	J1210	223.9	4.46	-3.78
J1204	7	0.73	0.31	J1210	226.1	4.70	-0.50
J1204	7.4	0.01	0.92	J1210	227.9	4.56	-0.63
J1204	8	1.05	1.33	J1210	230.1	3.24	0.73
J1204	8.3	0.58	1.00	J1210	232.1	3.35	-1.85
J1204	19.3	-2.78	-1.19	J1210	234.2	3.05	-0.43
J1204	20	-2.67	-1.22	J1210	236.1	3.36	-2.36
J1204	20.4	-4.02	-1.56	J1210	238.2	2.97	1.66
J1204	20.9	-3.77	-0.94	J1210	242.1	3.32	-3.67
J1204	21.1	-3.60	-0.62	J1210	244.6	3.30	-1.54
J1204	21.9	-7.00	-1.30	J1210	245.9	3.85	0.80
J1204	22.3	-2.76	-4.21	J1210	248.4	3.74	-0.49
J1204	22.9	-4.21	-1.87	J1210	249.8	3.31	1.06
J1204	23.3	-3.96	-2.01	J1210	252.6	4.37	2.88
J1204	23.9	-3.94	-2.07	J1210	258.2	3.61	1.85
J1204	24.4	-4.07	-2.22	J1210	260.1	3.77	2.85
J1204	25.4	-2.88	-5.85	J1210	262.4	3.56	1.91
J1204	25.9	-6.53	-2.16	J1210	263.9	3.90	-1.23
J1204	27.2	-5.64	-2.07	J1210	276.1	2.64	0.95
J1204	27.5	-5.50	-1.89	J1210	359.9	4.68	2.34
J1204	29.6	-3.26	-0.82	J1210	416.5	3.25	2.13
J1204	30.9	-2.97	-2.95	J1210	418.1	3.39	1.40
J1204	32.2	-2.37	-0.28	J1210	420.1	3.51	-2.98
				J1210	422.1	3.98	0.91
J1207	0.5	1.74	1.43	J1210	424.2	3.90	1.94
J1207	2.6	0.18	-0.88	J1210	426.3	3.84	-0.30
J1207	3.9	0.89	-1.35	J1210	427.8	4.51	-4.08

J1207	14.9	0.85	-2.99	J1210	428.2	4.34	-3.66
J1207	57.2	2.16	-0.90	J1210	430.1	4.33	-4.03
J1207	58.4	5.16	1.02	J1210	431.8	4.37	-5.61
J1207	59.2	4.27	0.40	J1210	434	4.38	-1.94
J1207	60.8	3.99	-0.24	J1210	435.8	3.80	-0.92
J1207	61.7	3.95	1.14	J1210	438.1	3.74	-2.22
J1207	66.1	3.14	-1.14	J1210	440.1	2.49	-0.41
J1207	68	5.49	1.81	J1210	443.8	2.37	-1.22
J1207	73.5	5.00	1.13	J1210	446.1	1.76	0.04
J1207	81.9	-0.24	1.77	J1210	447	2.76	-2.40
J1207	99	1.28	1.08	J1210	448.2	3.07	-1.13
J1207	104.7	0.82	0.52	J1210	518	1.63	-7.28
J1207	109.1	-0.65	-0.71	J1210	518.5	1.44	-8.28
J1207	114.7	0.61	0.08	J1210	519	1.57	-9.06
J1207	118.9	1.21	-3.75	J1210	519.5	0.69	-8.49
J1207	122.5	1.19	-7.66	J1210	521.8	0.49	-2.94
J1207	145.9	3.32	-4.86	J1210	524.3	-3.62	-3.07
J1207	151.8	3.70	-5.61	J1210	526.1	-3.67	-3.10
J1207	156.1	3.13	-2.65	J1210	527.9	-3.69	-2.79
J1207	161.2	3.81	-0.38	J1210	531	-0.45	-4.68
J1207	165.4	2.45	0.83	J1210	534	0.27	-4.94
J1207	171.2	3.55	-3.24				
J1207	176.4	3.63	-2.28	J1301	20	5.13	-5.34
J1207	179.9	4.14	1.66	J1301	22	5.00	-4.60
J1207	184.4	2.45	-0.80	J1301	24	4.70	-6.65
J1207	188	3.86	1.23	J1301	25.9	4.05	-3.00
J1207	189.7	3.84	-0.75	J1301	28.3	3.04	-0.31
J1207	196	3.29	-1.72	J1301	30.2	4.91	-2.75
J1207	201	3.17	-0.10	J1301	32.2	5.09	-3.11
				J1301	33.8	5.49	-0.38
J907	0	4.10	-1.69	J1301	36	5.06	-0.79
J907	1	2.18	-3.54	J1301	40.2	5.28	0.70
J907	23.5	3.60	-0.40	J1301	42.9	5.40	-0.44
J907	24.5	3.44	0.96	J1301	43.8	4.98	2.31
J907	25.5	3.29	0.22	J1301	48	4.14	1.91
J907	26.5	2.43	0.40	J1301	49.7	4.40	1.77
J907	27.5	-0.15	-4.55	J1301	54.5	4.00	0.52
J907	29.5	2.05	-0.31	J1301	56.4	4.41	1.94
J907	30.5	3.23	0.77	J1301	58.1	4.21	0.90
J907	31	3.32	1.81	J1301	60.2	4.62	1.76
J907	32	2.30	-0.81	J1301	62.6	4.10	2.42
J907	33	3.41	0.59	J1301	62.8	3.87	1.93
J907	34	2.75	0.05	J1301	65.6	1.61	-7.04
J907	35	2.62	-3.58	J1301	66.4	1.72	-5.92
J907	36	2.66	-3.56	J1301	68.6	1.45	-7.10
J907	37	3.13	-1.69	J1301	70.7	2.12	1.39
J907	38	3.04	-1.47	J1301	72.6	2.13	1.89
J907	39	3.59	0.33	J1301	74.7	1.32	0.70
J907	40	2.79	0.39	J1301	76.4	0.83	1.72
J907	41	2.64	0.27	J1301	100.5	1.42	-0.28
J907	42	2.40	0.03	J1301	102.7	0.27	-3.42
J907	43	2.10	0.52	J1301	104.5	0.89	-3.58
J907	44	2.08	-0.27	J1301	107.6	0.46	-0.26
J907	45	2.20	0.64	J1301	112	2.64	-2.67
J907	46	2.27	0.03	J1301	115	4.52	0.14
J907	47	2.30	-0.88	J1301	118	4.89	-0.87
J907	48	2.39	-0.57	J1301	120.8	4.32	-3.88
J907	49	2.54	0.12	J1301	123.7	4.05	-2.95

J907	50	2.77	-0.40	J1301	126.9	4.30	-3.20
J907	51	2.79	-0.24	J1301	130.4	3.92	-3.16
J907	52	2.04	-2.45	J1301	133	4.27	-2.59
J907	53	2.20	-0.12	J1301	137.1	4.47	-1.91
J907	54	3.51	1.11	J1301	140.3	4.39	-1.59
J907	55	2.22	-0.79	J1301	147.1	4.54	-1.31
J907	56	1.32	0.72	J1301	149.8	4.69	-1.26
J907	57	-0.04	1.15	J1301	152.6	4.80	-0.12
J907	58	-2.24	3.09	J1301	165.3	4.53	-0.45
J907	59	-0.91	1.92	J1301	168	4.85	0.04
J907	60	0.36	0.20	J1301	171	4.86	0.50
J907	61	-0.06	-0.60	J1301	177.8	3.99	-2.97
J907	62	1.42	-0.67	J1301	176.7	4.69	-0.19
J907	63	1.72	0.50	J1301	179.7	4.23	-2.03
J907	64	1.48	-2.60	J1301	185.2	4.24	-1.62
J907	65	2.22	-0.63	J1301	187.1	4.92	1.19
J907	66	0.96	-4.19	J1301	190	4.58	-0.40
J907	67	1.57	-0.03	J1301	193	5.01	1.36
J907	68	0.20	1.35	J1301	196.2	4.72	0.40
J907	69	-1.39	1.53	J1301	199.5	4.44	-2.03
J907	70	-2.12	2.08	J1301	202.3	4.52	-1.22
J907	71	-2.18	2.65	J1301	205.6	4.82	0.45
J907	72	-1.34	-0.03	J1301	208	4.51	0.22
J907	73	-2.08	2.54	J1301	211	4.11	-2.69
J907	74	-1.76	1.85	J1301	214.5	5.06	1.77
J907	75	-0.86	0.60	J1301	217.7	4.12	-2.23
J907	76	-1.33	-2.97	J1301	220.3	4.36	1.44
J907	77	-1.22	-1.27	J1301	223	5.01	1.96
J907	78	-1.77	1.66	J1301	226.1	5.13	3.22
J907	79	-0.99	-0.28	J1301	229.5	5.32	2.93
J907	80	-1.15	-0.09	J1301	232.1	5.08	1.04
J907	81	-2.90	2.27	J1301	235.4	4.89	0.26
J907	82	-3.28	2.00	J1301	238	4.99	1.48
J907	83	-1.24	-2.25	J1301	241	5.17	1.97
J907	84	-0.89	-0.53	J1301	243.1	4.92	0.66
J907	86	0.85	0.35	J1301	246	4.92	0.95
J907	88	0.54	-0.07	J1301	248.3	4.91	1.41
J907	90	2.34	-0.19	J1301	252	4.45	0.28
J907	92	2.58	2.34	J1301	255.1	4.70	1.09
J907	94	1.94	2.03	J1301	258.1	4.89	2.21
J907	96	1.82	2.56	J1301	261.2	4.75	1.72
J907	98	2.54	1.33	J1301	267	5.13	2.72
J907	100	1.71	-3.02	J1301	267	5.14	2.70
J907	102	2.63	0.50	J1301	270.1	4.83	1.08
J907	104	2.43	1.43	J1301	273	4.82	2.16
J907	106	2.77	1.75	J1301	275.1	4.77	0.65
J907	108	2.41	1.43	J1301	279.2	4.89	2.58
J907	110	2.30	1.62	J1301	282	4.87	2.72
J907	112	2.67	0.64	J1301	285.1	4.64	1.17
J907	114	2.89	1.01	J1301	288.2	5.03	1.63
J907	116	2.59	-0.13	J1301	291	4.99	1.46
J907	118	2.87	0.79	J1301	293	4.77	0.56
J907	120	3.08	2.04	J1301	296	4.95	2.30
J907	122	2.53	0.85	J1301	299.1	4.64	-0.31
J907	124	2.68	0.39	J1301	302.2	4.84	2.73
J907	126	2.94	1.26	J1301	305	4.44	-1.27
J907	128	2.73	-0.01	J1301	307.8	5.17	3.36
J907	130	2.88	0.51	J1301	310.1	4.47	0.87

J907	132	2.71	0.57	J1301	312.9	4.68	2.52
J907	134	3.08	1.38	J1301	316	4.32	1.76
J907	136	3.18	1.37	J1301	319	4.71	1.88
J907	138	3.14	0.66	J1301	322.3	4.43	2.16
J907	140	2.22	-1.09	J1301	325	4.08	1.34
J907	142	2.98	0.32	J1301	328.2	4.66	-0.16
J907	144	3.04	0.07	J1301	331.1	4.45	-0.10
J907	146	3.20	-0.55	J1301	334	4.94	1.16
J907	148	3.17	0.21	J1301	337	4.78	0.84
J907	150	3.44	0.48	J1301	340	4.06	-2.67
J907	152	3.01	-0.33	J1301	343.1	4.53	0.05
J907	154	2.58	-1.83	J1301	346	4.08	-0.25
J907	156	3.11	0.78	J1301	349.4	4.78	2.10
J907	158	3.25	0.21	J1301	352	4.49	1.46
J907	160	3.05	1.26	J1301	355.7	4.57	2.97
J907	162	3.36	-0.28	J1301	361.5	3.55	2.44
J907	164	3.31	0.24	J1301	364.6	3.77	0.25
J907	166	3.31	0.08	J1301	368.5	3.40	-1.92
J907	168	3.07	0.30	J1301	371.2	3.61	-1.02
J907	170	3.13	0.19	J1301	374.9	3.45	-0.49
J907	172	2.41	-0.21	J1301	377	3.66	1.25
J907	174	1.68	-2.25	J1301	380	3.87	0.85
J907	176	2.38	0.73	J1301	385.5	4.09	1.54
J907	178	0.94	1.73	J1301	387.5	3.88	0.16
J907	180	0.49	-1.47	J1301	389.7	3.44	-0.28
J907	182	2.03	1.87	J1301	391.2	3.22	2.19
J907	184	1.70	2.32	J1301	393	3.05	1.08
J907	186	1.95	0.89	J1301	394.7	3.93	0.38
J907	186	1.93	0.89	J1301	396.4	3.83	0.70
J907	188	1.22	0.79	J1301	397.4	4.05	0.63
J907	190	2.73	1.59	J1301	401.3	1.37	1.13
J907	192	2.47	0.70	J1301	401.7	1.71	0.45
J907	194	2.87	1.03	J1301	402.9	0.04	0.85
J907	194	2.83	1.10				
J907	196	2.62	0.19	J908	0.1	2.33	-4.60
J907	198	2.79	1.00	J908	1	3.95	-4.75
J907	200	2.69	0.94	J908	1.2	3.78	-4.79
J907	202	2.33	0.31	J908	3	2.71	-4.24
J907	204	1.64	-1.47	J908	3.8	3.44	-3.29
J907	205.5	-0.93	1.67	J908	4.4	2.09	-3.25
J907	206	0.44	-0.09	J908	23.6	2.91	-2.00
J907	208	0.43	1.51	J908	25.4	1.28	-0.76
J907	210	-0.08	3.13	J908	26	1.52	0.61
J907	212	0.06	1.51	J908	26.4	0.55	0.71
J907	214	0.40	0.58	J908	27	0.94	0.12
J907	216	-0.24	1.12	J908	28.8	1.26	-2.28
J907	218	0.02	1.17	J908	28.9	2.48	-0.69
J907	218	0.04	1.17	J908	30	3.19	0.95
J907	220	-0.49	2.23	J908	32	2.59	-0.49
J907	222	-0.28	2.07	J908	34	2.62	-0.06
J907	224	1.54	0.91	J908	36	2.93	0.84
J907	226	1.13	1.18	J908	38	2.48	-0.30
J907	228	1.59	0.63	J908	40	2.12	-0.10
J907	230	1.98	0.16	J908	46	0.90	-1.31
J907	232	1.62	1.98	J908	48	1.00	-1.23
J907	234	1.07	1.61	J908	50	1.07	-0.04
J907	236	1.51	1.18	J908	52	0.74	-4.05
J907	238	1.52	1.33	J908	54.1	-1.15	0.47

J907	240	1.68	0.73	J908	56.6	-2.16	-0.48
J907	242	2.14	1.49	J908	57.1	-3.25	0.21
J907	244	1.84	4.04	J908	58	-2.21	-0.72
J907	246	1.53	0.90	J908	60.4	-0.37	-3.31
J907	248	0.52	0.22	J908	61.5	2.76	0.03
J907	250	1.02	1.19	J908	62.5	2.85	-0.37
J907	252	1.07	2.43	J908	63.3	1.94	1.00
J907	254	1.22	2.38	J908	64.2	0.42	-0.83
J907	256	0.81	1.65	J908	64.6	2.48	-0.54
J907	258	0.75	1.54	J908	64.8	-0.80	-0.87
J907	260	0.80	2.11	J908	64.9	2.11	0.55
J907	262	1.00	2.05	J908	66.4	-0.19	0.16
J907	264	0.52	-0.49	J908	68.1	1.50	0.66
J907	266	0.09	1.99	J908	69.1	2.13	1.16
J907	268	0.35	0.75				
J907	270	0.68	0.83	J1122	0.1	1.82	-1.98
J907	272	0.09	2.56	J1122	2	2.05	-1.38
J907	274	1.90	2.02	J1122	3.9	0.59	-3.24
J907	276	1.95	1.45	J1122	6	1.60	-0.59
J907	278	3.21	2.82	J1122	8	1.48	-0.41
J907	280	2.69	2.74	J1122	12	0.12	-0.98
J907	282	2.66	3.15	J1122	14	-0.60	-1.30
J907	284	2.40	2.64	J1122	16	1.33	-3.50
J907	286	2.22	3.81	J1122	18	3.13	0.28
J907	288	3.31	4.25	J1122	20	2.86	-0.89
J907	290	1.33	1.73	J1122	22	1.50	-0.72
J907	292	1.10	2.04	J1122	24	1.41	-1.40
J907	294	0.63	3.28	J1122	26	1.38	-1.35
J907	296	2.89	3.71	J1122	28	1.53	-1.31
J907	298	3.02	5.16	J1122	30	1.13	-1.98
J907	300	1.57	3.38	J1122	32	0.84	-2.55
J907	302	1.56	2.82	J1122	34	1.68	-1.72
J907	304	3.09	3.74	J1122	36	-0.23	-2.52
J907	306	3.60	3.97	J1122	38	-0.15	-3.01
J907	308	1.24	3.27	J1122	40	1.37	-3.25
J907	310	-0.57	4.40	J1122	42	1.73	-2.02
J907	312	0.99	2.92	J1122	44	1.29	-1.78
J907	314	1.57	3.25	J1122	48	1.64	-1.45
J907	316	1.74	2.28	J1122	50	1.45	-1.84
J907	318	0.97	3.50	J1122	52	2.21	-0.99
J907	320	3.11	2.49	J1122	54	2.01	-1.39
J907	322	3.76	2.75	J1122	58	2.91	-1.35
J907	324	3.03	1.76	J1122	60	2.69	-1.39
J907	326	4.17	2.91	J1122	62	1.81	-1.66
J907	328	3.17	3.99	J1122	64	0.59	-0.94
J907	330	2.66	3.95	J1122	66	0.54	-0.74
J907	332	3.18	2.12	J1122	68	0.87	-1.96
J907	334	2.90	1.27	J1122	70	-0.05	-1.48
J907	336	1.81	3.69	J1122	73	0.14	-1.13
J907	338	2.30	4.29	J1122	75	0.08	-1.19
				J1122	77	0.53	-1.82
J1222	30.9	2.65	-3.27	J1122	79	0.26	-1.64
J1222	31.4	3.57	-2.92	J1122	83	0.43	-2.86
J1222	31.8	3.36	-2.55	J1122	85	1.61	-2.42
J1222	32.2	2.14	-3.47	J1122	87	1.62	-0.20
J1222	39.3	2.37	-4.50	J1122	89	1.89	-0.99
J1222	39.9	2.68	-4.20	J1122	91	1.16	-0.57
J1222	45.7	2.62	-4.48	J1122	93	0.03	-0.17

J1222	59.7	4.04	-0.87	J1122	95	-1.98	-2.17
J1222	60.6	3.51	-1.05	J1122	97	-5.61	-2.89
J1222	63.2	3.82	-0.22	J1122	99	-4.34	-1.87
J1222	62.1	3.39	0.14	J1122	101	-2.35	-4.36
J1222	64.3	3.37	-0.89	J1122	103	-0.96	-5.18
J1222	65.2	3.60	-0.23	J1122	105	-1.08	-3.85
J1222	65.7	3.08	0.48	J1122	107	-4.34	-1.84
J1222	66.1	-0.80	-5.76	J1122	114	-0.84	-3.99
J1222	66.2	0.21	-0.64	J1122	116	-0.27	-4.86
J1222	67	0.63	0.20	J1122	118	0.53	-4.62
J1222	69.1	2.11	0.00	J1122	120	0.98	-5.73
J1222	70.8	3.69	0.98	J1122	122	1.22	-4.46
J1222	72.9	2.84	-0.06	J1122	124	1.69	-2.57
J1222	74.9	3.83	0.22	J1122	126	2.36	-3.15
J1222	77.3	2.84	-2.04				
J1222	77.4	3.46	0.03	J1223	37.7	3.84	-1.43
J1222	81.1	3.13	-0.51	J1223	38.8	4.01	-0.34
J1222	83.1	3.23	-1.25	J1223	39.8	2.33	-0.66
J1222	85	3.02	-2.98	J1223	40.9	1.12	0.88
J1222	86.9	3.08	-2.09	J1223	43.2	-3.44	-4.00
J1222	89.6	2.81	-0.81	J1223	46.3	-2.70	-1.73
J1222	92.1	1.97	0.24	J1223	48.3	0.20	-0.66
J1222	94	1.49	-1.69	J1223	51.4	-2.84	2.51
J1222	96	1.66	-0.35	J1223	53.8	0.05	1.17
J1222	98.6	1.66	-0.38	J1223	56.3	1.17	-0.17
J1222	100	1.24	-1.16	J1223	59.1	1.16	1.43
J1222	104.3	1.22	-0.63	J1223	59.7	1.85	1.60
J1222	109.2	1.59	-1.81	J1223	62.3	1.80	-0.42
J1222	110.1	2.15	-1.36	J1223	65.1	1.63	0.91
J1222	111.3	2.00	-0.41	J1223	76	0.38	-1.73
J1222	112.7	1.90	-1.49	J1223	78.1	1.74	-0.62
J1222	113.3	2.44	-1.93	J1223	80	1.23	-1.09
J1222	116.2	-1.20	-0.47	J1223	82.2	2.63	-0.16
J1222	118	0.70	-0.62	J1223	84.3	1.37	-0.28
J1222	120.5	2.18	0.45	J1223	86.1	0.96	2.35
J1222	122	2.26	0.91	J1223	88.2	1.87	-1.03
J1222	123.5	2.61	1.67	J1223	90	0.97	-1.18
J1222	125.5	-2.86	1.17	J1223	92.2	1.31	2.13
J1222	126.9	0.54	-2.67	J1223	94.4	1.47	1.65
J1222	130.1	1.35	-4.17	J1223	96.9	1.30	0.75
J1222	130.2	2.17	-2.65	J1223	99	2.21	0.26
J1222	134	1.79	-4.04	J1223	101.1	2.24	-2.86
J1222	136.1	2.25	-2.27	J1223	105.6	0.73	-3.30
J1222	138.4	2.31	-2.13	J1223	107.2	0.90	-3.78
J1222	139.9	2.38	-1.16	J1223	110.1	1.34	-4.10
J1222	142	2.29	-0.71	J1223	112.1	0.45	-5.28
J1222	146.2	1.91	-1.20	J1223	94	0.35	-3.23
J1222	147.1	0.88	1.19	J1223	117.3	0.67	-1.91
J1222	150.4	1.40	0.19	J1223	121.1	-0.03	-1.72
J1222	153.3	1.90	-0.66	J1223	124.1	0.41	-2.90
J1222	156.5	1.88	-0.14	J1223	128.6	-0.57	-3.00
J1222	158.2	2.22	-1.46	J1223	129.8	-1.18	-3.66
J1222	160	2.49	-0.53	J1223	132.1	-0.25	-2.94
J1222	162.1	1.82	0.05	J1223	134.3	0.77	-3.85
J1222	164.2	2.28	-0.25	J1223	136	2.00	-0.96
J1222	166	2.14	-1.28	J1223	136	1.94	-0.94
J1222	168	2.09	-1.75	J1223	137.5	0.43	0.81
J1222	169.3	2.57	-0.94	J1223	139.5	1.68	-0.90

J1222	172.8	2.23	-1.95	J1223	141.6	0.60	-0.33
J1222	175.1	1.32	-3.70	J1223	143.5	-0.11	1.65
J1222	177	0.37	-2.40				
J1222	178.5	-0.42	-4.29	J1019	27.7	0.53	-8.62
J1222	179	-0.22	-3.63	J1019	29.4	1.48	-7.34
J1222	191.5	-0.06	0.07	J1019	30.0	1.73	-6.77
J1222	194.5	-0.03	-1.57	J1019	31.4	-0.88	-11.37
J1222	196.2	-0.06	-0.98	J1019	37.3	0.36	-5.38
J1222	198	0.01	-0.63	J1019	37.7	0.84	-3.28
J1222	200	-0.20	-1.98	J1019	38.3	0.65	-6.59
J1222	202	-0.25	-1.56	J1019	39.1	0.44	-8.31
J1222	205	0.27	-0.39	J1019	39.6	0.98	-7.17
J1222	207.2	0.32	0.59	J1019	40.0	1.63	-5.02
J1222	208.1	1.23	1.97	J1019	40.5	1.20	-4.40
J1222	211.2	0.70	1.95	J1019	41.0	0.25	-2.28
J1222	212.9	0.27	1.50	J1019	79.6	4.25	-0.24
J1222	214.8	0.25	2.26	J1019	81.0	1.54	0.68
J1222	217.1	-0.04	0.81	J1019	82.0	4.06	1.34
J1222	218.9	-1.10	-0.65	J1019	83.0	4.55	-0.09
J1222	219.8	1.04	-2.03	J1019	84.0	4.76	1.37
J1222	221.2	0.74	2.52	J1019	85.0	4.76	1.13
J1222	225.1	-1.84	-0.21	J1019	86.0	4.70	-0.75
J1222	227.4	-0.21	-0.45	J1019	87.0	4.74	-1.65
J1222	229.9	-0.33	-3.86	J1019	88.0	4.99	-0.20
J1222	234.3	-1.63	-4.23	J1019	89.0	4.72	-0.67
J1222	232.1	-1.32	-4.47	J1019	90.0	4.83	0.82
J1222	235.4	-1.22	-3.79	J1019	91.0	4.46	0.65
J1222	237.4	-0.79	-2.80	J1019	92.0	2.74	-0.99
J1222	239.6	-1.62	-1.42	J1019	96.0	2.74	-3.21
J1222	240.6	-1.48	-4.50	J1019	97.0	2.84	-3.78
J1222	242.7	-1.80	-3.27	J1019	98.0	2.53	-4.97
J1222	248.6	-1.06	-2.93	J1019	99.0	3.08	-3.25
J1222	250.7	-0.54	-3.22	J1019	100.0	2.99	-3.42
J1222	253.7	-0.67	-4.03	J1019	101.0	3.01	-3.05
J1222	255.5	-0.82	-2.72	J1019	103.0	2.93	-3.41
J1222	257.6	-0.26	-3.35	J1019	104.0	2.88	-3.20
J1222	259.8	-0.79	-2.29	J1019	105.0	2.49	-2.87
J1222	261.9	-0.77	-1.94	J1019	106.0	3.21	-5.55
J1222	264.9	-0.78	-1.51	J1019	107.0	2.61	-5.02
J1222	269	-0.05	-3.14	J1019	109.0	3.47	-4.26
J1222	271.2	-0.70	-2.66	J1019	110.0	3.35	-3.85
J1222	275.1	-0.03	-3.64	J1019	112.5	2.92	-3.10
J1222	280.6	-0.64	-1.54	J1019	116.0	2.95	-3.26
J1222	282.4	-1.13	-4.22	J1019	117.0	2.92	-3.34
J1222	286.7	-0.72	-3.47	J1019	118.0	2.91	-3.08
J1222	287.8	-0.56	-3.12	J1019	119.0	3.30	-2.52
J1222	290.2	-0.28	-3.88	J1019	120.0	3.09	-2.33
J1222	293.3	-0.52	-5.90	J1019	122.0	2.07	-2.80
J1222	294.4	0.14	-5.66	J1019	123.0	2.53	-2.50
J1222	296.5	0.20	-4.00	J1019	124.0	2.95	-5.60
J1222	304.2	-1.37	-2.31	J1019	127.0	3.12	-5.72
J1222	306.7	-1.50	-2.19	J1019	128.0	2.50	-3.83
J1222	311.2	0.26	-2.83	J1019	129.4	2.11	-6.80
J1222	314	1.95	-1.06	J1019	130.4	1.45	-3.30
J1222	316.3	-0.37	-3.26	J1019	131.0	2.75	-5.44
J1222	321.6	0.38	1.17	J1019	132.0	1.39	-3.22
J1222	323.7	1.81	2.02	J1019	133.0	0.97	-2.60
J1222	325.9	0.35	1.33	J1019	134.0	1.36	-2.62

J1222	327.5	1.94	1.89	J1019	135.0	1.80	-4.04
J1222	330.6	2.09	3.84	J1019	137.0	1.32	-2.31
J1222	332.5	1.85	1.43	J1019	140.0	4.14	1.98
J1222	335	3.20	3.08	J1019	141.0	3.51	0.66
J1222	337.3	3.05	1.18	J1019	142.0	3.41	1.62
J1222	339.2	1.80	2.66	J1019	145.0	1.09	0.06
J1222	343.4	3.80	3.40	J1019	147.0	1.25	-3.83
J1222	346.2	3.11	3.80	J1019	151.0	1.62	-0.22
J1222	349.3	3.39	2.77	J1019	156.0	3.26	0.48
J1222	352.3	1.47	1.80	J1019	157.0	2.31	-0.29
J1222	354.4	2.64	2.14	J1019	158.0	3.18	-1.25
J1222	360.5	1.72	1.75	J1019	159.0	3.20	-1.04
J1222	362.6	1.86	0.63	J1019	163.0	2.48	-0.35
J1222	365.3	3.63	3.71	J1019	164.0	3.40	0.18
J1222	367.6	3.06	3.75	J1019	165.0	2.07	-0.05
J1222	372.7	3.01	1.77	J1019	173.0	3.64	-0.48
J1222	374.5	2.46	2.53	J1019	175.0	3.18	0.53
J1222	378.4	3.27	1.66	J1019	185.0	2.38	1.81
J1222	381.4	2.19	0.46	J1019	190.0	1.98	1.16
J1222	383.4	1.42	2.72	J1019	192.0	-0.92	1.74
J1222	385.6	3.11	0.38	J1019	194.0	-0.04	-0.07
J1222	388.5	2.70	0.43	J1019	196.0	0.22	-0.45
J1222	392.1	2.67	3.03	J1019	198.0	1.57	-0.92
J1222	394.5	2.26	-1.42	J1019	200.0	0.85	-1.34
J1222	396.5	2.96	1.80	J1019	202.0	1.54	-1.88
J1222	398.5	0.94	-1.13	J1019	204.0	1.77	-1.81
J1222	400.6	3.58	1.27	J1019	206.0	2.14	-1.79
J1222	404.4	2.87	1.90	J1019	208.0	2.29	-0.60
J1222	406.5	2.43	2.58	J1019	210.0	2.17	-1.39
J1222	408.5	2.81	1.82	J1019	212.0	0.94	-1.63
J1222	410.6	2.78	1.93	J1019	214.0	2.22	-1.33
J1222	412.4	2.54	2.07	J1019	216.0	0.05	1.17
J1222	412.9	1.01	-0.65	J1019	218.0	-0.52	-0.91
J1222	414.8	1.52	-1.12	J1019	220.0	-2.23	0.78
				J1019	222.0	0.26	-0.90
F928	6	3.19	-0.83	J1019	224.0	1.23	-2.68
F928	8	3.68	0.34	J1019	226.0	-0.14	-0.13
F928	10	3.66	0.34	J1019	228.0	-0.10	-0.26
F928	12	3.33	-0.66	J1019	232.0	0.21	-0.58
F928	14	3.28	-1.58	J1019	234.0	1.54	-6.80
F928	16	3.16	-1.74	J1019	236.0	1.17	-7.37
F928	18	3.55	-1.17	J1019	238.0	0.28	-4.88
F928	20	3.90	-0.67	J1019	240.0	0.28	-1.01
F928	22	3.29	-1.84	J1019	242.0	-2.35	-2.85
F928	24	3.13	-1.92	J1019	244.0	1.01	-3.94
F928	28	3.52	-1.20	J1019	248.0	2.55	-4.28
F928	28	3.52	-1.76	J1019	250.0	1.81	-2.91
F928	30	3.75	-0.82	J1019	252.0	1.90	-6.21

F928	32	3.18	-3.28	J1019	254.0	2.40	-8.18
F928	34	3.50	-0.85	J1019	258.0	3.34	-4.46
F928	36	3.66	1.69	J1019	260.0	2.99	0.89
F928	38	3.26	-0.74	J1019	262.0	4.03	-1.71
F928	40	3.91	-0.13	J1019	264.0	4.21	-3.16
F928	42	3.27	0.85	J1019	266.0	5.77	-2.13
F928	46	3.05	-1.21	J1019	268.0	2.85	-6.77
F928	50	2.81	-2.12	J1019	269.0	2.62	-6.77
F928	54	2.85	-3.56	J1019	275.0	2.76	-4.62
F928	58	3.16	-2.03	J1019	279.0	1.81	-5.17
F928	62	3.07	-2.02	J1019	283.0	2.63	-4.79
F928	66	2.78	-0.05	J1019	287.0	1.96	-4.42
F928	70	3.17	-1.44	J1019	291.0	2.74	-4.65
F928	74	3.05	0.62	J1019	295.0	1.84	-4.66
F928	78	3.81	-0.86	J1019	299.0	2.62	-1.20
F928	82	3.25	1.73	J1019	303.0	-0.27	0.79
F928	88	4.27	0.77	J1019	307.0	1.35	-1.01
F928	92	3.65	-0.69	J1019	311.0	0.82	-2.02
F928	96	3.71	2.32	J1019	315.0	2.79	0.45
F928	100	4.75	2.35	J1019	323.0	1.89	-0.50
F928	104	4.39	2.23	J1019	327.0	1.83	-5.17
F928	108	4.92	2.73	J1019	331.0	3.30	1.82
F928	112	4.98	-0.49	J1019	333.0	2.88	1.69
F928	116	4.32	1.27	J1019	337.0	3.58	1.86
F928	120	4.48	1.08	J1019	341.0	3.36	0.94
F928	124	3.94	2.66	J1019	345.0	2.89	1.30
F928	128	5.06	1.47	J1019	355.0	4.03	1.82
F928	130	4.84	3.18	J1019	365.0	4.00	0.03
F928	136	5.10	3.00	J1019	375.0	3.99	2.25
F928	140	4.79	-0.35	J1019	385.0	3.43	-1.80
F928	144	3.95	0.88	J1019	395.0	2.88	0.69
F928	148	4.51	0.28	J1019	405.0	2.97	0.75
F928	152	3.91	1.36	J1019	415.0	3.19	-0.57
F928	156	4.35	3.13	J1019	425.0	3.72	-0.30
F928	160	4.79	-1.48	J1019	429.0	3.59	-0.47
F928	164	3.65	2.89	J1019	433.0	3.70	0.11
F928	168	4.93	2.97	J1019	437.0	3.78	0.30
F928	172	4.66	2.43	J1019	441.0	3.63	0.79
F928	176	4.54	2.55	J1019	447.0	4.46	1.30
F928	180	4.63	2.62	J1019	451.0	4.08	-0.43
F928	184	4.78	1.04	J1019	455.0	3.91	0.92
F928	188	3.75	1.40	J1019	459.0	4.14	2.63
F928	192	4.57	2.33	J1019	463.0	3.95	0.64
F928	196	4.80	0.11	J1019	465.0	3.47	0.09
F928	200	3.96	-1.80	J1019	467.0	3.61	0.01
F928	204	4.00	1.18	J1019	471.0	3.75	1.87
F928	208	4.61	1.09				
F928	212	4.65	0.44				
F928	216	4.19	2.95				
F928	220	4.54	1.88				
F928	228	4.30	1.26				
F928	232	3.70	0.92				
F928	236	3.60	2.12				
F928	240	3.50	0.49				
F928	244	3.47	2.05				
F928	248	3.44	0.94				
F928	256	3.23	2.43				
F928	260	3.88	2.31				

TABLE 1. SUMMARY OF CALLISON LAKE FORMATION LITHOFACIES (PAGE 1)

Lithofacies	Composition	Bedding Style/Structures	Depositional Environment	Distribution
Siliciclastic Facies:				
<u>F1: Pebble to granule conglomerate</u>	Clast- to matrix-supported conglomerate; light grey to brown; dominated by quartz and chert with occasional lithics; well-rounded to subangular; moderate sorting; hematite and silica cement	Thin- to thick-bedded; occasionally forming distinct transgressive lags; faint ripple- to dune-scale trough and tabular cross-bedding up to 20 cm thick; erosional base, lenticular geometry, and fining-up packages common; associated with F2 and F3	Braided fluvial channels. Tidal/Estuarine? Amalgamated channel deposits with abrupt lateral facies change; difficult to discern distinct point-bar or lateral accretion geometries; locally reworking underlying strata; transgressive lag	Restricted to HM
<u>F2: Sandstone</u>	Quartz and chert arenite and wacke; very fine- to coarse-grained with occasional granules; moderate sorting; well-rounded to subangular; abundant hematite staining; tan yellow to brown; silica cement	Thin- to medium-bedded; commonly with unidirectional and symmetrical ripple cross-bedding and parallel lamination; occasionally amalgamated; mud chip intraclasts common; interbedded with F3 and forming fining-up packages with F1	Coastal plain/Estuarine. Maybe local shoreface? Evidence for floodplain or tidal flat deposition with clear tidal influence in the form of distinct tidal ravinement surfaces and fining-up packages	Restricted to HM
<u>F3: Variegated siltstone and shale</u>	Variegated shale (red, yellow, green, and purple) interbedded with siltstone; locally contains diagenetic dolostone lenses and iron formation; heavily silicified in certain horizons; occasional very-fine sand sized particles	Locally abundant mudcracks and scours; mostly planar-laminated and associated with F1 and F2; occasional coarsening-upward packages with ripple cross lamination; flaser-bedding and ball and pillows locally	Floodplain or peritidal mud flat/lagoon. Generally associated with transition into marginal marine inner ramp to lagoonal deposits; interbedded sandstone units could represent crevasse splays in a fluvial setting	Restricted to HM
<u>F4: Black shale</u>	Black to dark grey shale; TOC to 3.5 wt. %; occasionally silicified with abundant chert nodules; locally contains vase-shaped microfossils; minor silt and very-fine sand; Fe-oxides, pyrite, sphalerite, ankerite(?) locally	Planar-laminated and fissile; occasionally silicified and more resistant; locally interbedded with various lithofacies and no evidence for wave or storm activity; poor sorting with coarser grained horizons	Outer-inner ramp subtidal or lagoonal. Suspension depositor in a low-energy setting; elevated TOC tied to episodic restriction?	HM, TRM
Evaporite/Talc Facies:				
<u>F5: Talc-rich shale</u>	Black talc-rich shale; commonly silicified with abundant chert nodules; organic-rich with TOC up to 4 wt%	Planar-laminated to nodular bedded; abundant diagenetic chert; occasional soft-sediment deformation; intimate association with F6	Restricted lagoonal to sahbka. Playa lake? Suspension deposition in a low-energy setting; evidence for seismic or storm-generated disruption	TM, TRM?
<u>F6: Interbedded talc and dolostone</u>	Black talc-rich shale interbedded with microbial and sucrosic light to dark dolostone; commonly silicified with chert nodules	Nodular- to planar-bedded with microbial lamination; teepees structures, mudcracks, and evaporite pseudomorphs after anhydrite and gypsum common; horizons of talc-shale chip intraclast conglomerate common; local overturned stromatolites and seismites	Sahbka to peritidal mud flat. Evidence for intertidal to peritidal deposition; episodic restriction with sulfate deposition; wave-, storm-, and seismically-generated structures common; complex paragenetic history with multiple Mg-silicate and evaporite transformations	Restricted to TM
Carbonate Facies:				
<u>F7: Stromatolitic bioherm/biostrome</u>	Orange-yellow to dark grey (fresh) stromatolitic doloboundstone; minor terrigenous silt and frosted quartz grains; occasional vase-shaped microfossils in disseminated organic matter; microbial sheaths after cyanobacteria	Meter- to decimeter-thick stromatolitic biohermal/biostromal buildups interbedded with F3,4; stromatolites range from domal to columnar with both low- and high-inheritance forms; micritic, oolitic, and intraclastic fill; isolated to laterally linked forms	Middle- to inner-ramp subtidal or lagoonal. Isolated to linked stromatolitic patch reefs or buildups associated with a subtidal depositional setting; possibly lagoonal with no evidence for exposure or wave activity	HM, TRM

HM – Heterolithic member, TM – Talc member, PM – Platform member, TRM – Transitional member

TABLE 1. SUMMARY OF CALLISON LAKE FORMATION LITHOFACIES (PAGE 2)

Lithofacies	Composition	Bedding Style/Structures	Depositional Environment	Distribution
<u>F8: Stromatolitic doloboundstone</u>	Light to dark grey stromatolitic doloboundstone; occasional black chert and ferruginous clay-rich laminae; close association with stromatolite-rich intraclast wackestone and grainstone	Centimeter- to meter-thick stromatolitic structures; forms range from laterally-linked, low relief, and domal to high-inheritance columnar structures; intercolumnal fill micritic, oolitic, pisolitic, and intraclasts; scours and corrosive structures common	Middle-inner ramp subtidal. Heterogeneous assemblage of stromatolitic morphologies; common wave-generated scours and association with F9 and F10; meter-scale high relief domes suggest occasional deeper water setting	All members.
<u>F9: Microbialite</u>	Light to dark grey microbial doloboundstone; occasional black chert nodules, terrigenous silt, fine- to medium-sized frosted quartz grains; distinguished from F8 by flat, crinkly lamination	Centimeter- to meter-thick microbialite; mm-scale undulatory and wavy lamination; locally containing fenestrae (birds' eye), teepee structures, and mudcracks; occasional seismically-disrupted and folded laminae; discrete intervals of intraclast conglomerate	Inner ramp intertidal to peritidal. Intertidal deposition evidenced by intimate association with F8 and wave- or storm-generated intraclasts; exposure surfaces common; discrete parasequences with F6,8, and 10	All members.
<u>F10: Intraclast grainstone/wackestone</u>	Light to dark grey dolomitic intraclast grainstone and wackestone; almost exclusively composed of clasts from F8 and F9; minor rudstone	Thin- to thick-bedded, tabular to subrounded, sand- to cobble-sized intraclastic grainstone and wackestone matrix-supported with scoured and erosive bases, occasional crude lamination, and faint grading; commonly randomly oriented	Inner ramp subtidal. Subtidal deposits associated with high-energy storm- or wave-generated events that scour and rework material from F9 and F10; no evidence for subaerial exposure	PM, TRM
<u>F11: Dolograinstone</u>	Light to dark grey dolograinstone composed of ooids, peloids, pisoids, and/or oncoids; minor terrigenous silt component	Thin- to thick-bedded and massive to finely laminated; occasional trough cross-bedding and planar laminated but generally quite massive and crudely stratified; intimately associated with F8, F9, and F10; occasional low-angle cross-bedding	Inner ramp subtidal to intertidal bar or shoal complex. High-energy subtidal deposition associated with tidal sand bars and/or migrating barrier complex; 3D morphology not worked out but commonly interbedded with diverse lithofacies	Restricted to PM
<u>F12: Thin-bedded dolomicrite/siltite</u>	Dark grey to black dolomicrite and dolosiltite; commonly composed of microcrystalline and neomorphosed dolospar and dolomicrite; occasional hints of peloidal precursor allochems; ferruginous clay drapes and black chert common	Thin-bedded and laminated with mm-scale parallel lamination; occasional erosional base; black chert commonly late diagenetic and fabric destructive; no evidence for grading or turbidite deposition; commonly resemble Phanerozoic ribbon-bedded limestone	Mid-outer ramp subtidal. Generally constrained to transgressive horizons associated with relatively low-energy deposition; ferruginous drapes could indicate hardground conditions or suspension rainout during highstand; occasionally deformed during seismic activity	PM, TRM
Diagenetic Facies: <u>F13: Recrystallized dolostone</u>	Light grey to white sucrosic dolostone with fabric-destructive diagenetic recrystallization; occasional oolitic ghosts	Medium- to thick-bedded and massive; crude stratification and commonly impossible to tell precursor lithology; abundant secondary isopachous, drusy, and botryoidal cements; veins common; locally contains clotty secondary fabric that resembles "thrombolitic" texture	Unclear but commonly associated with subaerial exposure surfaces and coarser-grained lithofacies.	All members
<u>F14: Breccia</u>	Grey to white dolostone breccia; commonly silicified with angular clasts of dolostone; matrix composed of both terrigenous silt, secondary dolomite spar, and dolosiltite; clasts range from sand- to boulder-sized; local well-rounded quartz sand within cavity fill	Massive and generally thick-bedded; crude stratification with clear horizons indicating subaerial exposure and karst development; clasts composed exclusively of underlying strata in puzzle-fitting fabric; local grykes and irregular cavities filled with terrigenous silt and sand	Inner ramp intertidal to peritidal. Clearly associated with subaerial exposure and karst development in dolostone lithologies; possibly associated with uplift and fault breccia in certain localities	PM, TRM

HM – Heterolithic member, TM – Talc member, PM – Platform member, TRM – Transitional member