

**School of Earth and Planetary Science**

**The Geologic Record of Two Proterozoic Supercontinents within the  
Belt Basin Region of Western North America**

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**This thesis is presented for the Degree of  
Doctor of Philosophy  
of  
Curtin University**

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## **Declaration**

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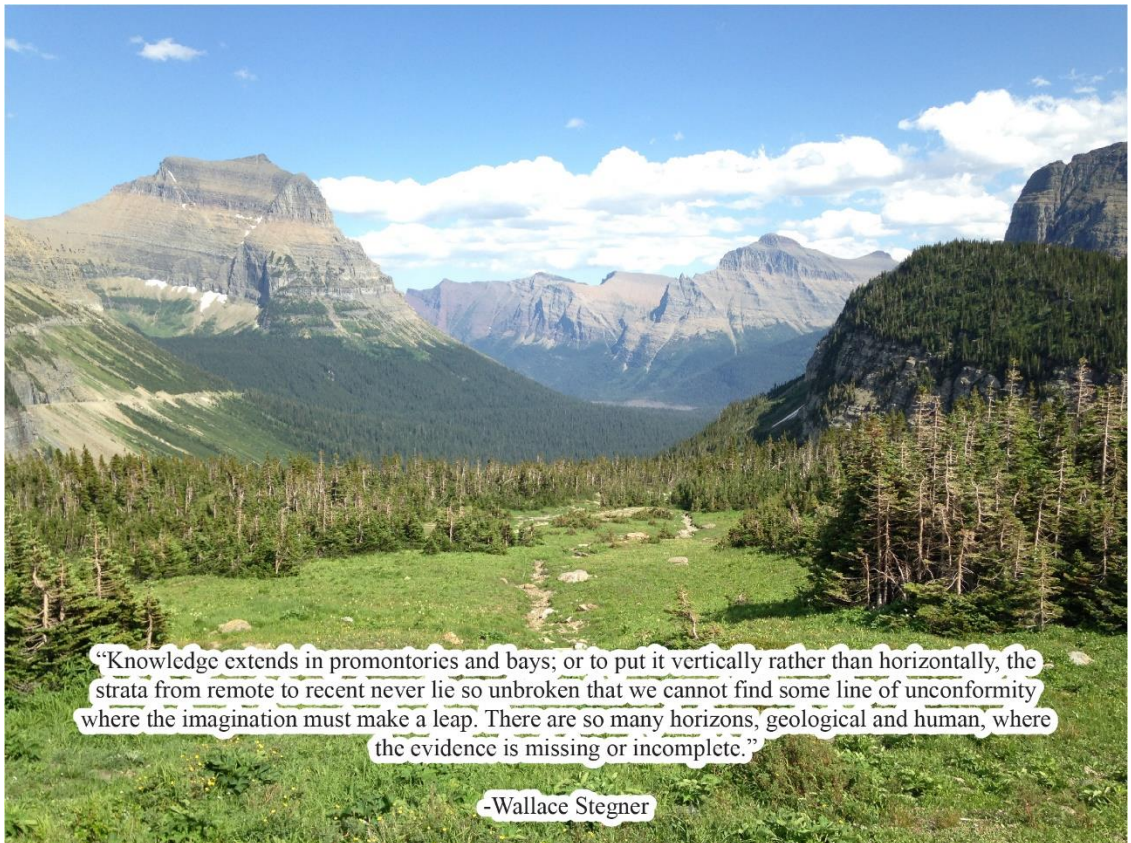
## **Acknowledgment of Country**

I acknowledge that during my time at Curtin University, I lived and worked on the traditional lands of the Wadjuk people of the Nyungar Nation. The fieldwork conducted and geologic samples collected for this scientific inquiry are from the traditional lands of the Kalispel, Coeur d'Alene, and Nez Perce people. I wish to pay my deepest respects to the ancestors and community members, past, present, and emerging of these First Nation peoples. While I strove to honor the geologic significance of the North American study region, I acknowledge and appreciate that the value and meaning of these lands surpass metrics measurable by scientific inquiry.

## **Abstract**

The cyclical amalgamation and dispersal of most of Earth's continents, a process known as the supercontinent cycle, has been operable for at least the last 1.8 billion years. Many consider the supercontinent cycle to be the first-order control on the non-uniform spatial and temporal distribution of Earth's resources, and to have played a governing role in the environmental and biologic evolutionary changes that facilitated life. Early attempts at reconstructing Precambrian supercontinents often relied upon simplified correlations of singular geologic entities such as paired rift margins. However, more recent attempts to reconstruct ancient supercontinent configurations require multidisciplinary studies considering an even broader set of geologic entities such as basement provinces, orogenic histories, basin histories, paired rift margins, plume events, global plate kinematics, and paleomagnetic data. As a result, reconstructions have become more detailed, but the compounding effect of regional tectonic disagreements is still evident in wildly variable proposed global paleogeographic configurations. The Belt Basin region of western Laurentia has long been recognized to contain a key, yet debated, geologic record of the timing and configurations of the first supercontinent Nuna (1.6–1.3 Ga; also called Columbia), and its successor Rodinia (0.9–0.7 Ga). This study addresses some of the most significant of the Proterozoic magmatic, stratigraphic and metamorphic tectonic debates in this region, including A) the origin of localized North American Magmatic Gap magmatism, B) the tectonostratigraphic framework of Proterozoic strata, and the occurrence, timing, source, and implications of allegedly non-Laurentian derived detrital zircon within these strata, and C) the tectonic implications of enigmatic ca. 1.3–1.0 Ga metamorphism reported in this region. By bolstering field observations with U/Pb geochronology (zircon, monazite, apatite), additional zircon isotopic data (Lu/Hf, O-), and metapelite phase equilibria modelling, the results of this study strengthened commonly advocated ties between southeastern Australia and the Belt Basin region (USA) within supercontinent Nuna. However, the results also identified discrepancies between the Belt Basin region and commonly favored conjugate blocks (southeastern Australia or South China) within popular "SWEAT" or "Missing-Link" Rodinia configurations. Consequently, this work supports the accuracy of recently proposed Nuna configurations but advocates the need for more work on Rodinia reconstructions.





“Knowledge extends in promontories and bays; or to put it vertically rather than horizontally, the strata from remote to recent never lie so unbroken that we cannot find some line of unconformity where the imagination must make a leap. There are so many horizons, geological and human, where the evidence is missing or incomplete.”

-Wallace Stegner

## **Author Acknowledgments**

This document marks the culmination of a long personal and academic journey that spanned several relationships, continents, decades, and too many crappy student housing rentals. To paraphrase E. A., at times this journey did feel crooked, winding, lonesome, and dangerous...but man, did those hard times sure lead to some amazing views and fulfilling moments! I have gratitude for so many people who helped me find my path along the way. I would have been lost alone.

The first and perhaps the most important guides I had along the way were my parents. Mom and Dad, thank you. From encouraging my curiosity as a kid to answering those 3 A.M. stress-induced sleeplessness phone calls, your love, and unwavering reassurance that I am defined by far more than my failures or accolades did not go unappreciated. I could not have done this without you. To my sisters, cousins, aunts, and other family members... thank you for your support. However, it was perhaps your unassuming general disinterest in geology that always seemed to remind me when it was needed most that geology is not who I am, it is just something I enjoy doing.

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## Publications during the course of this thesis

### Peer-reviewed journal articles:

**Brennan, D.T.**, Link, P.K., Li, Z.X., Martin, L., Johnson, T., Evans, N.J., and Li, J., 2022, Closing the “North American Magmatic” Gap: Crustal evolution of the Clearwater Block from multi-isotope and trace element zircon data: *Precambrian Research*, v. 369, p. 106533, doi:10.1016/j.precamres.2021.106533.

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**Brennan, D.T.**, Brian Mahoney, J., Li, Z.X., Link, P.K., Evans, N.J., and Johnson, T.E., 2021, Detrital zircon U–Pb and Hf signatures of Paleo-Mesoproterozoic strata in the Priest River region, northwestern USA: A record of Laurentia assembly and Nuna tenure: *Precambrian Research*, v. 367, p. 106445, doi:10.1016/j.precamres.2021.106445.

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**Brennan, D.T.**, Li, Z., Rankenburg, K., Evans, N., Link, P.K., Nordsvan, A.R., Kirkland, C.L., Mahoney, J.B., Johnson, T., and McDonald, B.J., 2021, Recalibrating Rodinian rifting in the northwestern United States: *Geology*, v. 49, no 6., 617–622, doi:10.1130/G48435.1

**Brennan, D.T.**, Mitchell, R.N., Spencer, C.J., Murphy, J.B., and Li, Z., 2021, A tectonic model for the Transcontinental Arch: Progressive migration of a Laurentian drainage divide during the Neoproterozoic-Cambrian Sauk Transgression: *Terra Nova*, p. 1–11, doi:10.1111/ter.12528.

### Conference presentations:

**Brennan, D.T.**, Link, P.K., Li, Z.X., Martin, L., Johnson, T., Evans, N.J., and Li, J., 2022, Closing the “North American Magmatic” Gap: Crustal evolution of the Clearwater Block from multi-isotope and trace element zircon data, Geological Society of Australia Earth Sciences Student Symposium – Western Australia, Perth, 25 November 2021.

*Awarded Best Talk of Conference*

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# Chapter 1: Introduction

## 1.1. The Supercontinent Cycle

One of the paradoxes of living on a tectonically active planet is that the same geologic processes responsible for exposing ancient rocks making them available for study also commonly obscures their original nature (Fig. 1.1). Consequently, there is regularly a correlation between the ambiguity in our understanding of tectonic events and their antiquity. During the late 20<sup>th</sup> and into the 21<sup>st</sup> century, the development of analytical methods to measure a variety of geochemical and isotopic systems within rocks and minerals, revolutionized the field of geology (e.g., Holmes, 1946; Goldschmidt, 1954; Stacey and Kramers, 1975; Frost et al., 2001; Hawkesworth and Kemp, 2006). Not only do geochemical and isotopic methods facilitate absolute dating of geologic and tectonic events, they can also be utilized as tracers/recorders of geologic and tectonic processes in situations where the original context has been obscured by younger events (e.g., Faure, 1977). When combined with field methods, the advent of geochemical and isotopic analytical methods has opened entirely new frontiers in the study of Earth systems.



**Fig. 1.1:** Panoramic photograph (looking northeast) of Belt Supergroup strata in Glacier National Park, Montana USA. The striking exposures of these Mesoproterozoic rocks results mostly from Late Cretaceous thrusting and uplift, followed by Pleistocene glacial incision. While these processes exposed these rocks for study, they also contributed to regional debates over particular stratigraphic correlations for decades. Photography by Brennan in 2018.

Prior to the availability of robust isotopic analytical methods, limited outcrop exposure, extensive tectonic over-printing, and the lack of absolute chronologic constraints made interpreting the geologic record of Precambrian rock assemblages particularly troublesome. As such, our knowledge of the 88% of earth's history that constitutes the Precambrian has only recently begun to be comparable with our

knowledge of younger Phanerozoic events. However, 20<sup>th</sup> century advancements in our understanding of geochemical and isotopic systems within rocks and minerals have gradually revealed that Earth's geosphere has an ancient and geologically active past.

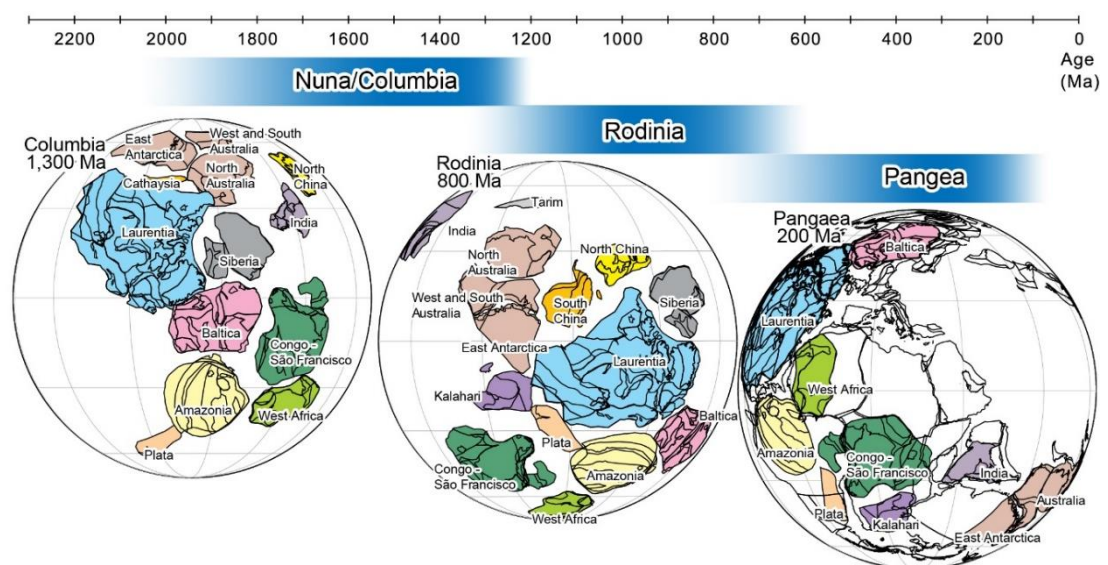
Congruent with the proliferation of geochemical and isotopic studies was a growing scientific acceptance of plate tectonic theory (e.g., McKenzie and Parker, 1967; Dewey and Horsfield, 1970). Not only was this theory consistent with modern geologic phenomena, most notably it also offered a testable mechanism for the movement of earth's continents allowing evaluation of Wegner's (1912) assertion that most of Earth's continents may have once been assembled into a singular landmass or "supercontinent." Wegner (1912) called this landmass Pangea (derived from the ancient Greek 'pan' meaning all, and 'gaea' meaning land). It didn't take long before others proposed that similar geologic processes likely occurred deeper, in the pre-Pangea, geologic past.

In 1970, Valentine and Moores were the first to officially propose the occurrence of a pre-Pangea, late Precambrian supercontinent subsequently named Rodinia (from the Russian word 'rodit' meaning 'to beget' or 'to give birth'; McMenamin and McMenamin, 1990). The origin of this name is particularly suitable, as it was within the nutrient rich extensive shallow seas that resulted from globally widespread rifting associated with the breakup of Rodinia, that the proliferation of lifeforms during the Neoproterozoic–Cambrian "explosion" took place (e.g., Marshall, 2006; Peters and Gaines, 2012; Brocks et al., 2017; Hoffman et al., 2017). The concept and configuration of this late Precambrian "Rodinia" supercontinent has been long debated since the early 1990's (e.g., Moores, 1991; Dalziel, 1991; Hoffman, 1991). These disagreements are often based on conflicting correlations of ca. 1300–1000 Ma "Grenville-age" orogens along which the supercontinent assembled and ca. 800–600 Ma rift margins that record subsequent supercontinent breakup.

In the late 1990's and early 2000's, the recognition of widespread ca. 2100–1800 Ma orogens led to the suggestion of the occurrence of an even older, pre-Rodinia supercontinent called "Nuna" (an Inuktitut term meaning "the land" or "all land", Hoffman, 1997) and/or "Columbia" (Rogers and Santosh, 2002; Zhao et al., 2002). As expected, the configuration of this Paleo-Mesoproterozoic supercontinent has also attracted much attention and debate within the geologic community. In particular, many geologic models favour similar configurations for both Rodinia and Nuna (e.g., Cawood and Hawkesworth, 2014; Li et al., 2019 and references within). This has

raised discussions on whether they should be considered fully separate individual supercontinents (e.g., Condie et al., 2002) and consequently highlights the relatively poor understanding of the breakup and ensuing assembly processes of Earth's allegedly first and second supercontinents.

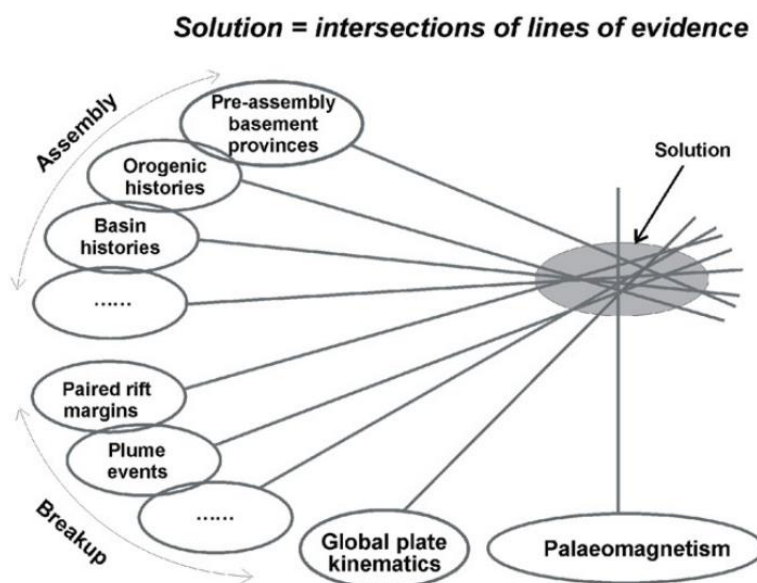
Despite intensive debate on the details, it has nonetheless generally been accepted that the assembly and subsequent dispersal of most of Earth's continents into supercontinents was a recurrent process for at least the past 1.8 billion years (Fig. 1.2; Condie, 2002; Nance et al., 2014; Evans et al., 2016; Li et al., 2019; Mitchell et al., 2021). The widespread recognition of long-term global cyclicality in tectonic processes is perhaps one of the most important developments in our understanding of the Earth since the acceptance of plate tectonic theory itself. Although it is now known that the supercontinent cycle was one of the first-order controls on the unequal spatial and temporal distribution of earth resources (e.g. Cawood and Hawkesworth, 2014), the drivers and mechanisms of tectonic cyclicality associated with these processes have many unanswered questions. For example, some argue that mantle-convection patterns are the predominate control on the style and evolution of Earth's tectonic cyclicality (Li et al., 2019), while others advocate that surficial erosion driven changes in subduction patterns are the primary first order control (Sobolev and Brown, 2019). An important first step in addressing these ongoing disagreements is the establishment of reliable, accurate and testable paleogeographic tectonic framework reconstructions throughout Earth's history.



**Fig. 1.2:** Timeline of supercontinents through time with paleogeographic reconstructions of Nuna at ca. 1300 Ma (also called Columbia), Rodinia at ca. 800 Ma, and Pangea at ca. 200 Ma (from Mitchell et al., 2021).

## 1.2. A multidisciplinary approach to evaluating Precambrian supercontinent tectonics

Early attempts at Precambrian global paleogeographic tectonic reconstructions often relied heavily on correlation of singular geologic entities such as paired rift margins (e.g., Eisbacher, 1985). However, as Precambrian datasets grew, the correlation of multiple geologic entities such as paired rift margins and paleomagnetic data (Bell and Jefferson, 1987); and paired rift margins, paleomagnetic data, and orogenic histories (Dalziel, 1991; Moores, 1991; Hoffman, 1991) quickly became commonplace. More recently, it is generally accepted that robust paleogeographic reconstructions must be consistent with even broader set of geologic entities such as basement provinces, orogenic histories, basin histories, paired rift margins, plume events, global plate kinematic models, and paleomagnetic data (Fig. 1.3; e.g., Li et al., 2008). As such, the increasing consideration of multiple complementary datasets has aided in the generation of more detailed (e.g., Merdith et al., 2021) and hopefully more accurate reconstructions. However, these more detailed reconstructions have also highlighted how even minor disagreements and/or varying interpretations of what some may consider regional geologic discrepancies can be compounded into significant disagreements between global paleogeographic tectonic reconstructions. For example, disagreements on the Neoproterozoic geologic history of South China has led it to be placed centrally (Li et al., 2008), on the periphery (Wang et al., 2020), or perhaps even completely removed from Rodinia (Park et al., 2021).



**Fig 1.3:** A schematic diagram illustrating the importance of a multidisciplinary approach to reconstructing and evaluating palaeogeography (from Li et al., 2008). Note how when only a single line of geologic evidence is considered, there are often many non-unique solutions.

Consequently, regions where Precambrian basement, orogenic, basin, rift, and/or plume geologic records are present in close proximity often warrant detailed study, as robust tectonic models of these regions are a prerequisite for accurate and testable global paleogeographic tectonic models. Comparison of the tectonic records of these regions allows them to be linked in the past with other currently far afield regions containing similar histories, thus acting as geologic “piercing points.” As a result, solving global-scale tectonic and paleogeographic disagreements is impossible as long as notable regional geologic debates persist.

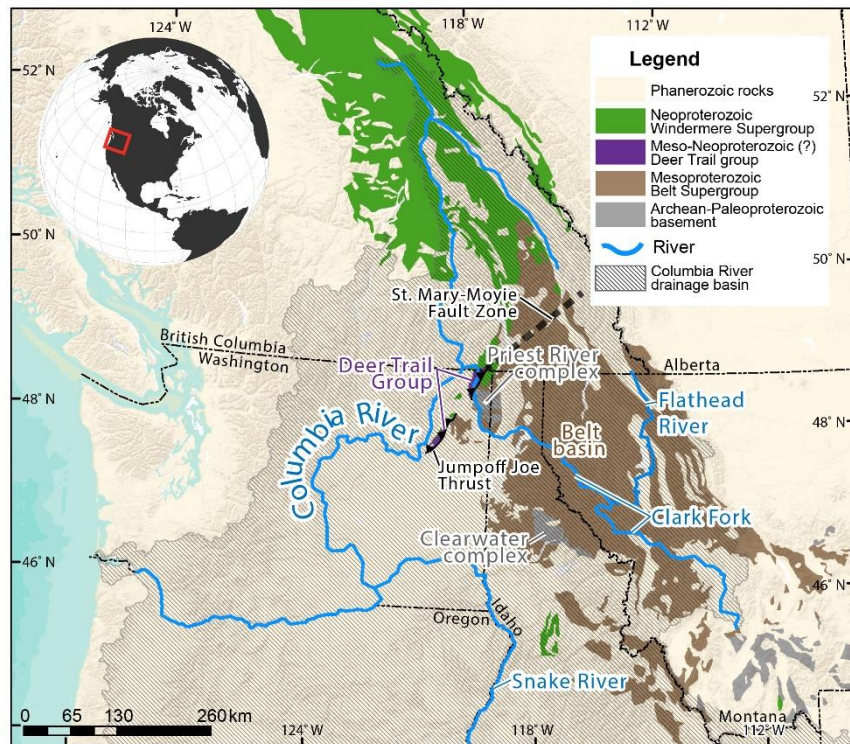
### **1.3. The Belt basin region – A critical Nuna and Rodina piercing point**

As discussed, “Nuna” is a term in Inuktitut (the language of the indigenous people of the North American Arctic region) meaning “the land,” or “all land,” and Hoffman (1997) deserves credit for its first use to describe the Paleoproterozoic assembly of Laurentia and Baltica. Zhao et al. (2002) presented an extensive overview of global ca. 2100–1800 Ma orogens, and demonstrated that most other major Archean cratonic blocks such as those of North China, Australia, Siberia, India, South Africa, West Africa, and South America assembled along ca. 2000–1800 Ma orogens, suggesting synchronicity with Laurentia and Baltica’s assembly. Following Rogers and Santosh (2002), Zhao et al. (2002) correlate these widespread ca. 2100–1800 Ma collisional orogens to reflect the assembly of a Paleo-Mesoproterozoic supercontinent they call Columbia. The Columbia name was first proposed by Rogers and Santosh (2002) who state that “the key evidence for its existence is the relationship between eastern India and the Columbia region of North America” (pg. 5). However, within the same Gondwana research volume, Meert (2002) presents the first paleomagnetic constraints for the Columbia Paleoproterozoic supercontinent, and in slight disagreement with the ca. 2100–1800 Ma assembly of Columbia advocated by Rogers and Santosh (2002), Meert notes that Paleomagnetic constraints suggest supercontinent assembly occurred primarily between ca. 1770–1500 Ma.

Consequently, an enduring controversy over the past two decades is not only whether the “Nuna” or “Columbia” name is better suited for the potential supercontinent that assembled in the late Paleoproterozoic (e.g., Meert, 2012), but if this supercontinent assembled coevally with widespread ca. 2000–1800 Ma “Trans-



Hudson” age orogens (e.g., Zhao et al., 2002), or if final assembly occurred later, probably after ca. 1800 Ma but prior to ca. 1500 Ma (e.g., Meert, 2002). In regards to the first (nomenclature) controversy, various arguments have been made recently to support both names. Harrison and St.-Onge (2022) note that the meaning of “Nuna” being “the land,” or “all land,” (in Inuktitut) is a more appropriate supercontinent name as it is akin to the meaning of Pangea (derived from the ancient Greek ‘pan’ meaning all, and ‘gaea’ meaning land). A sentiment shared by this author, and reflected by the preferred usage throughout this work. While others (e.g., Wang et al., 2021) suggest that the name Columbia, based off the “Columbia” region of western North America (namely Washington State) is more appropriate as it was the name used by the authors that first attempted global-scale reconstructions of this purported supercontinent (e.g., Rogers and Santosh, 2002; Zhao et al., 2002). While, naming preferences between myself and Rogers and Santosh (2002) differ, I do concur that the “Columbia” (which essentially comprises the western portion of Belt basin) region of western North American is critically understudied considering its long-recognized importance for global scale paleogeographic reconstructions (Fig. 1.4).



**Fig. 1.4:** Simplified geologic map of the northwestern United States, and southwestern Canada. The main Proterozoic tectonostratigraphic packages of this region, within the Columbia River drainage basin, consists of the extensive Mesoproterozoic Belt Supergroup (shown in brown), and Neoproterozoic Windermere Supergroup strata (and its correlatives, shown in green) which have both be integral in development of early Rodinia and Nuna paleographic reconstructions (e.g., Bell and Jefferson, 1982; Rogers and Santosh, 2002).

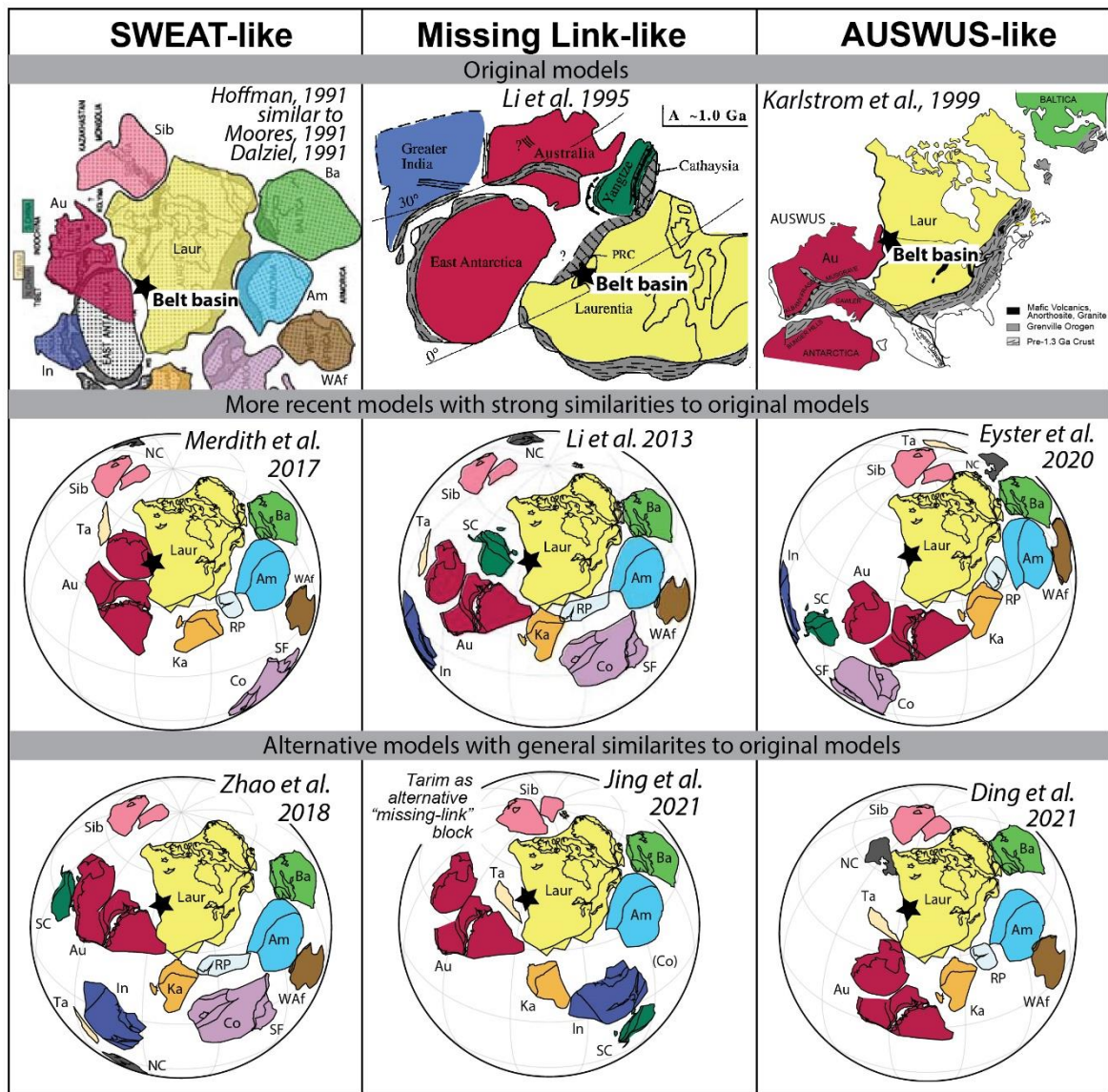


In fact, a growing body of paleomagnetic, geochronologic, and sedimentary provenance data has made the original “eastern India and the Columbia region of North America” correlation of Rogers and Santosh (2002) fall out of favor. Instead most recent models suggests links between the eastern margin of the North Australian craton (NAC) and the northwestern margin of Laurentia during assembly of supercontinent Nuna based on shared provenance, metamorphic and magmatic histories (Thorkelson et al., 2001; Ross and Villeneuve, 2003; Pisarevksy et al., 2014; Nordsvan et al., 2018; Pourteau et al., 2018; Volante et al., 2020, 2022). A lesser constrained southwest Laurentia-Mawson continent (Gawler Craton of South Australia and the Terre Adelie Craton of East Antarctica; Payne et al., 2009) connection is also suggested for this time (Mulder et al., 2015; Verbaas et al., 2018; Goodge et al., 2017). Based on similarities between crustal provinces in the Cathaysia Block of South China and southern Laurentia, Li et al. (1995), Yao et al. (2017) and others instead have proposed that the Cathaysia Block was between western Laurentia and Australia during Nuna time.

However, paleomagnetic constraints have demonstrated that the proposed Australia-Laurentia connection within Nuna could not have persisted uninterrupted during evolution to the subsequent supercontinent, Rodinia. Paleomagnetic constraints advocate that Australia and Laurentia separated between ca. 1300 and 1200 Ma (Pisarevsky et al., 2003; Kirscher et al., 2020), which suggests the occurrence of a complete Wilson cycle along the western Laurentia margin during the Nuna to Rodinia transition (Betts and Giles, 2006). However, a regional geologic record of this purported event is historically poorly recognized in western Laurentia.

The ensuing configuration of continents west of Laurentia within the supercontinent Rodinia (at ca. 1000 Ma) is an enduring (and perhaps endearing) controversy. Several original models are over 20 years old and continue to be revised and favoured by recent publications without a clear scientific consensus (see Fig. 1.5). Some of these leading models include a southwest U.S.–East Antarctic connection (SWEAT; Moores, 1991, Hoffman, 1991; Dalziel, 1991; Merdith et al., 2017; Zhao et al., 2018), possibly with a Rodinia-forming Grenville-age (Sibao) orogeny between (the Laurentia-connected) Cathayasia and the incoming Yangtze blocks of South China (Missing-Link; Li et al., 1995; Li et al., 2013), or with Australia located along the southwestern U.S. (AUSWUS; Karlstrom, 1999; Eyester et al., 2020) or Mexico (AUSMEX; Wingate et al., 2002). Completely different western Laurentia conjugates such as Tarim (Wen et al., 2017; 2018; Jing et al., 2021), or the North China Craton

(Ding et al., 2021) have also been suggested. However, at least one early model, the Siberian Connection (Sears and Price, 2000), which suggests that Siberia was conjugate to western Laurentia within Rodinia has generally fallen out of favor due to its inconsistency with most paleomagnetic data (Li et al., 2008; Pisarevsky et al., 2021; Evans, 2021).



**Fig. 1.5:** Variable Rodinia reconstruction models including the original models (top row), more recently proposed “updated” interpretations that share strong similarities with the original models, and alternative recently proposed models that have general similarities with original models. The location of the Belt basin region, the focus of this study area, is indicated by the black star in each reconstruction. Note that in relation to the pertinent regions of this study, most of these models differ in the location of Australia/Antarctica (in dark red) relative to western Laurentia, and the occurrence (or not) of small intervening blocks between Laurentia and Australia/Antarctica. Figure adapted from Evans (2021 and references within). Abbreviations include: Am, Amazonia; Au, Australian cratons including Antarctic Mawsonland; Ba, Baltica; Co, Congo; In, India; Ka, Kalahari; Laur, Laurentia; NC, North China; RP, Rio Plata; SC, South China; SF, Saño Francisco; Sib, Siberia; Ta, Tarim; Waf, West African craton.

## 1.4. Regional debates with global implications

Several unique and debated geologic records within the Belt basin of western Laurentia make this region a distinctive Laurentian piercing point for Precambrian paleogeography (Fig. 1.6). Addressing these regional debates will allow for the evaluation of many of the competing global models presented above, and thus have the potential to significantly influence our understanding of Proterozoic supercontinent processes.

Within the Priest River and Clearwater metamorphic complexes (Fig. 1.4) within the Belt Basin region, 2670 to 2650 Ma Neoproterozoic and 1880 to 1840 Ma Paleoproterozoic magmatic rocks of the Clearwater block are exposed (Vervoort et al., 2016). Interestingly, the ca. 1480–1380 Ma Belt Supergroup strata that overlie these basement rocks have long been known to lack detrital zircon ages consistent with this Laurentian basement and instead contain abundant ca. 1700–1500 Ma detrital zircon grains (Ross and Villeneuve, 2003; Lewis et al., 2010). Many of these grains have ages that fall within the ca. 1610–1490 Ma “North American Magmatic Gap” (NAMG), a time interval during which magmatism was rare in Laurentia, but common in Australia. Consequently, this relationship is often used to justify the proximal relationship of Australia with western Laurentia during the Mesoproterozoic (e.g., Ross and Villeneuve, 2003). However, the presence of NAMG-age magmatism in South China (Xu et al., 2019) has also brought into question the uniqueness of this interpretation.

In the Priest River complex (Fig. 1.4), a thrust bounded ca. 1580 Ma granite (Evans and Fischer, 1986) represents one of the only ca. 1610–1490 Ma magmatic rocks in Laurentia. Its occurrence may negate the necessity of any non-Laurentian sources for the NAMG-age zircon grains in the Belt Supergroup. Within the Priest River complex, a thin coarse quartzite unit is either the basal unit of the Belt Supergroup, or is unconformably overlain by lower Belt Supergroup strata. Limited geochronology results suggest this unit lacks these unique ca. 1610–1490 Ma detrital zircon grains (Doughty et al., 1998).

Most of the metamorphism along western Laurentia is usually attributed to the late Mesozoic and younger Cordilleran orogeny (Dickinson, 2004). However, in the Clearwater complex, metasedimentary Belt Supergroup strata contain an enigmatic spread of pre-Cordilleran ca. 1380–1000 Ma Lu/Hf garnet ages (Zirakparvar et al., 2010; Nesheim et al., 2012). The older range of these garnets may correspond with ca.

1380–1350 Ma bimodal plutonism within the Belt Basin. This magmatism could be associated with the East Kootenay “orogeny” (McMechan and Price, 1982; Evans and Zartman, 1990; Doughty and Chamberlain, 1996; McFarlane and Pattison, 2000). The significance of the East Kootenay orogeny is contentious, but instead of an actual orogeny, it may represent a period of renewed subsidence along the western margin of Laurentia (in the Belt-Purcell basin) and potentially reflect the rifting of Australia from western Laurentia (Doughty and Chamberlain, 1996).

The tectonic context of the younger “Grenville” ca. 1150–1000 Ma metamorphic ages (Zirakparvar et al., 2010; Nesheim et al., 2012) are also debated. Some interpret a static thermal disturbance in the Belt region driven by magmatism at depth during this time interval (Doughty and Chamberlain, 2008). However, others advocate that the ca. 1100 Ma garnet growth requires crustal thickening (Zirakparvar et al., 2010) suggesting the presence of a Grenville-age orogen along western Laurentia during Rodinia assembly. Several global models propose either a ca. 1100 orogeny correlatable with the orogeny in the potentially adjacent South China (Sibao orogeny in the Missing-Link model, Li et al., 2008b) or a significant dextral transpressional margin (Wen et al., 2008, Mulder et al., 2018b) along western Laurentia that could account for this debated western Grenville-age orogenic event.

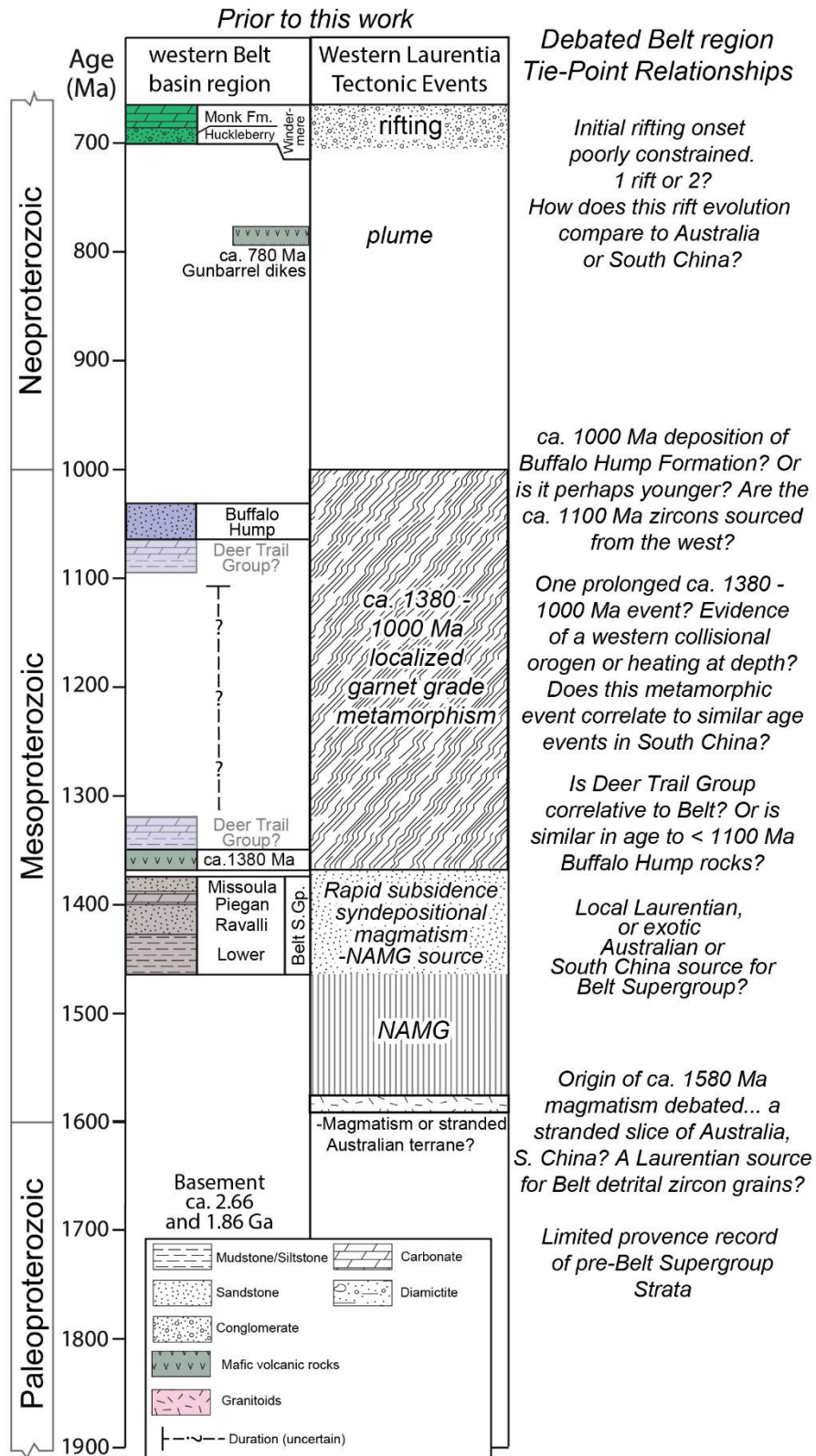
Along the northwestern edge of the Belt Basin, just west of the Priest River complex, the Buffalo Hump Formation and underlying Deer Trail Group strata crop out (Fig. 1.4). These rocks comprise the western-most exposures of Mesoproterozoic (and possibly Neoproterozoic) rocks in Laurentia. The entire Buffalo Hump Formation and Deer Trail Group were initially correlated with ca. 1470–1380 Ma Belt Supergroup to the east and across a significant thrust fault (the Jumpoff Joe Thrust), based on broad overall lithological similarities and the fact that they are overlain by the < 720 Ma Windermere Supergroup (Evans, 1987; Miller and Whipple, 1989). However, limited U-Pb analysis of detrital zircons from the Buffalo Hump Formation suggests a maximum depositional age of ca. 1100 Ma (Ross et al., 1992), indicating that it is too young to be part of the ca. 1380 Ma and older Belt Supergroup. This relationship requires either a significant unconformity between the Buffalo Hump Formation and the underlying allegedly Belt Supergroup correlative Deer Trail Group, or that the entire Deer Trail Group, which lacked any geochronologic constraints at the onset of this work, has also been miscorrelated.

The identification of ca. 1100 Ma detrital zircon grains within the Buffalo Hump Formation, and the relative absence of ca. 1100 Ma magmatism in western Laurentia, led Ross et al., (1992) to speculate that either: (a) The zircon grains were far-travelled or recycled, or (b) the grains record the former presence of Grenvillian orogenic activity within western Laurentia. However, the absence of a widespread Grenvillian-orogenic belt in western Laurentia, and subsequent evidence for far-travelled Laurentia-sourced ca. 1100 Ma detrital zircon grains in northern and western Laurentia (e.g., Rainbird et al., 1992; Yonkee et al., 2014) tends to support the first hypothesis of Ross et al., (1992). Nevertheless, others (e.g., Yao et al., 2017) have continued to advocate that the necessity of a proximal, non-Laurentian source of ca. 1100 Ma zircon grains for the Buffalo Hump Formation requires South China, and its ca. 1100 Ma Sibao orogeny, adjacent to the Belt Basin region within Rodinia.

In the 30 years since Ross et al., (1992) published the first detrital zircon results from the western Belt Basin region, ca. 1100 Ma detrital zircon grains have been found to actually be quite common in the mostly Cryogenian and Ediacaran Windermere Supergroup and correlative strata along the western margin of Laurentia (e.g., Gehrels et al., 1995; Yonkee et al., 2014; Matthews et al., 2018; Box et al., 2020). Windermere Supergroup strata are generally younger than ca. 720 Ma, and record the onset of widespread rifting along western Laurentia associated with the breakup of Rodinia (e.g., Ross, 1991). Several occurrences of ca. 780–720 Ma, pre-Windermere Supergroup, strata within localized basins (Chuar, Pahrump, and Uinta groups) may record localized earlier rifting (e.g., Dehler et al., 2010). However, so far these ca. 780–720 Ma basins are only identified south of the Belt basin region, which has led to proposals (e.g. Haldari et al., 2021) that the Belt basin segment of the margin, specifically the St. Mary-Moyie fault zone (Fig. 1.4), reflects a notable Neoproterozoic boundary along the rift margin.

The St. Mary-Moyie fault zone (Fig. 1.4) roughly aligns with a basement structure called the Vulcan zone and was likely active during Mesoproterozoic Belt Supergroup deposition (e.g., McFarlane and Pattison, 2000). A segment of the St. Mary-Moyie Fault zone may also be reactivated as the Mesozoic thrust fault that currently separates exposures of Deer Trail Group strata, from Belt Supergroup rocks to the east (Miller and Whipple, 1989). Consequently, there are disagreements on if the exact timing, style, and number of Neoproterozoic rifting events differed for the segments south and north of the St. Mary Moyie structure, which has significant

implications for Rodinia breakup processes in this region (e.g., Colpron et al., 2002; Lund et al., 2003; Hadlari et al., 2021).



**Fig. 1.6:** Time-space diagram for the key debated Paleo–Neoproterozoic geologic and tectonic “tie-point” events of the western Belt basin region (northeastern Washington and northern Idaho, USA), prior to the onset of this work in 2018.

## 1.5. Research Objectives

The main objective of this work is to provide insights into several regional tectonic debates in the Proterozoic geologic history of the northwestern United States, notably along the western extent of the Mesoproterozoic Belt Basin (Fig. 1.6). Recalibrating the geologic record of this important Laurentian “piercing point” region allows evaluation of debated Nuna and Rodinia supercontinent tectonic reconstructions. Broken down into individual objectives, this task was accomplished through:

- 1) Investigation of the origin of ca. 1.58 Ga North American Magmatic Gap Magmatism (NAMG) within the Clearwater Block, and further characterization of the isotopic composition of the nearby Archean–Proterozoic magmatic rocks. These results provided insights into the Proterozoic tectonic history of the region, and the provenance of overlying sedimentary units.
- 2) Geochronology and metamorphic study to resolve the timing, conditions, and origin of Meso–Neoproterozoic (Grenville-age) metamorphic events previously reported in the Clearwater region, and evaluate their relationship to similar events reported elsewhere along the western extent of the Belt Basin, and in potentially conjugate blocks.
- 3) Determination of the age, stratigraphic framework, and provenance of the Paleoproterozoic to Neoproterozoic strata of northeast Washington and northern Idaho (USA). Link these sedimentary rocks to potential source regions either within and/or outside Laurentia and determine the likely tectonic mechanisms for their deposition and provenance signatures.

Finally, these findings were integrated with the available Laurentian and non-Laurentian data to evaluate Nuna and Rodinia models focusing on the evolution of the western Laurentian margin in comparison to proposed conjugate regions, such as southeastern Australia, East Antarctica and South China during this time interval.



## 1.6. Thesis Structure

This thesis is presented as a series of four published papers and a manuscript prepared for submission to a peer-reviewed journal. After peer review and publication, a data processing error was identified that affected the publications presented in Chapters 2 and 3. Corrigendum's to correct both publications were issued and Chapters 2 and 3 have been updated to reflect this correction and consequently vary slightly from their original published form. The text and figures of the published works (Chapters 2, 3, 5 and 6) and work currently in preparation for submission (Chapter 4) are reproduced in full and have been reformatted for consistency in the thesis. Thus, there is some unavoidable repetition between these chapters, particularly regarding the geological background and methods sections as each chapter is intended to read as standalone document.

**Chapter 2: Closing the “North American Magmatic” Gap: Crustal evolution of the Clearwater Block from multi-isotope and trace element zircon data** presents new U/Pb, Lu/Hf, O-, and trace element data from ca. 2.67 Ga, 1.86 Ga, and 1.58 Ga “North American Magmatic Gap” igneous rocks within the Clearwater Block. These data provide insights into the magmatic record of western Laurentian during the assembly of supercontinent Nuna. This data allows assessment of prior speculations that North American Magmatic Gap magmatism in the Clearwater Block may be associated with long recognized similar-age magmatism in the Gawler Craton of South Australia, or more recently identified similar-age magmatism in East Antarctica or the Hainan Island of South China.

**Chapter 3: Detrital zircon U-Pb and Hf signatures of Paleo-Mesoproterozoic strata in the Priest River region, northwestern USA: A record of Laurentia assembly and Nuna Tenure** investigates the provenance of the Paleoproterozoic Gold Cup Quartzite, Mesoproterozoic Belt Supergroup, and likely late Mesoproterozoic Deer Trail Group in the northwestern USA. This work provides the first laser-ablation split stream U/Pb and Lu/Hf data from detrital zircon grains within these sequences, and provides insights into the source and stratigraphic extent of North American Magmatic Gap detrital zircon grains recognized within these sequences.



**Chapter 4: Grenville-age metamorphism within the Belt basin of western Laurentia** examines the cryptic late Mesoproterozoic metamorphism reported within the Clearwater Complex. This work provides the first Pressure-Temperature-time constraints for this event based on U/Pb monazite and apatite geochronology, and thermocalc thermodynamic modelling.

**Chapter 5: Recalibrating Rodinian rifting in the northwestern United States** presents new detrital zircon U/Pb and Lu/Hf data from the Buffalo Hump Formation. The Buffalo Hump Formation was previously alleged to have a ca. 1.0 Ga depositional age. However, utilizing a rapid laser-ablation analysis methodology, a minor but significant ca. 760 Ma detrital zircon population was identified. This revised depositional age constraint has important ramifications for the occurrence of ca. 1.0 Ga (Grenvillian) tectonism in western Laurentia and the timing of Rodinian rifting.

**Chapter 6: A tectonic model for the Transcontinental Arch: Progressive migration of a Laurentian drainage divide during the Neoproterozoic–Cambrian Sauk Transgression** provides a new perspective on tectonic provenance models for ca. 1.1 Ga detrital zircon grains in western Laurentia. This new tectonic model is supported by two-dimensional quantitative comparison of a compilation of published U/Pb and Lu/Hf datasets from Neoproterozoic to early Paleozoic strata in western Laurentia, and potential eastern/southeastern Laurentian igneous sources.

**Chapter 7: Synthesis and conclusions: The role of northwestern Laurentia in understanding global supercontinent processes.** In this chapter the refined Proterozoic history of the Belt basin region of western Laurentia is synthesized, and the consistency of global Proterozoic supercontinent models with the regional geologic record of this key revised piercing point is assessed. In particular, this chapter summarizes how this work revises our understanding of:

- 1) The Paleoproterozoic history of western Laurentia and how it relates to Nuna assembly, in particular was there evidence of Nuna assembly by ca. 1.85 Ga, or was it a later ca. 1.65 Ga process?

- 2) The Mesoproterozoic history of western Laurentia and how it relates to the Nuna to Rodinia transition, specifically what evidence is there for a ca. 1.1 Ga “western Grenville” orogenic event?
- 3) The Neoproterozoic history of western Laurentia and how it relates to Rodinia breakup, in particular what is the timing and style of rifting in western Laurentia? How does it relate to other suggested conjugate blocks?

In answering these questions, Chapter 7 brings the work presented in the individual chapters together and presents a revised Proterozoic tectonic model for the Belt basin region of western Laurentia. This regional tectonic model is then used to evaluate the feasibility of inherently more speculative global tectonic reconstructions, and the implications are discussed. As always, all scientific knowledge is conjectural, and while this work answers some questions, it also raises many more. Consequently, an important part of this concluding chapter is to emphasize uncertainties with the tectonic models resulting from this work and to highlight questions for future investigation.

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# **Chapter 2: Closing the “North American Magmatic” Gap: Crustal evolution of the Clearwater Block from multi-isotope and trace element zircon data**

## **Abstract**

Along the west-central margin of Laurentia, within the Priest River and Clearwater complexes, rare exposures of crystalline basement rocks of the Clearwater Block include “North American Magmatic Gap” (NAMG, ca. 1.61–1.49 Ga) ages. Elsewhere in this region, crystalline basement rocks are buried beneath thick deposits of the overlying Mesoproterozoic Belt Supergroup and younger sequences. The unique combination of magmatic basement ages and the detrital zircon components (which also include NAMG detrital ages) within the overlying Mesoproterozoic Belt Supergroup strata, has led researchers to identify the Clearwater Block region as a key tie-point for Proterozoic paleogeographic reconstructions. Some researchers even speculated that Proterozoic supercontinent events stranded exotic (possibly Australia, Antarctica or South China associated) basement terranes within the Clearwater Block. However, no comprehensive multi-isotopic data exist on the Neoproterozoic to Mesoproterozoic Clearwater Block crystalline basement leaving many of these speculations untested. We report new U–Pb, Lu–Hf and O isotopic data and trace element results of zircon grains for these rocks. Collectively, along with existing data these new results indicate a crustal evolution for the Clearwater Block that consists of: 1) ca. 2.67 Ga juvenile mantle-derived crustal growth as evident by mantle-like  $\delta^{18}\text{O}$  and slightly supra-chondritic (+4 to 0)  $\varepsilon_{\text{Hf}_t}$  values, 2) ca. 1.87 to 1.83 Ga melting and metamorphism of this ca. 2.67 Ga crust as recorded by ca. 1.86 Ga samples containing zircon with low (<0.1) Th/U ratios and retention of mantle-like  $\delta^{18}\text{O}$  values, and similar-age ca. 1.87–1.83 Ga samples containing zircon with higher (>0.1) Th/U ratios, and 3) ca. 1.58 Ga “NAMG” magmatism that records mantle-like  $\delta^{18}\text{O}$  but sub-chondritic (-5.5 to -9.5)  $\varepsilon_{\text{Hf}_t}$  values. We interpret the ca. 1.58 Ga NAMG magmatism to be consistent with (perhaps plume-driven) reworking of the Clearwater Block’s ca. 2.67 Ga lower crustal reservoir. Consequently, these results support a Laurentian origin for NAMG-age magmatism within the Clearwater Block, but confirm the

necessity of non-Laurentian sources for juvenile (+ $\epsilon$ Hf<sub>i</sub>) NAMG-age detrital zircon grains in the overlying ca. 1.47 Ga lower Belt Supergroup strata, thus providing important constraints for western Laurentia's conjugate during the early Mesoproterozoic.

## **2.1. Introduction**

The North American Magmatic Gap (NAMG) is an apparent tectono-magmatic lull from ca. 1.61 to 1.49 Ga within Laurentia (Van Schmus et al., 1993). This period of relative magmatic quiescence occurred after the culmination of the Paleoproterozoic Trans-Hudson (ca. 2.0–1.8 Ga) and Yavapai-Mazatzal-Mojave orogens (ca. 1.65 Ga), and prior to the start of extensive anorogenic magmatism within the Transcontinental Granite-Rhyolite Province at ca. 1.48 Ga (Ross and Villeneuve, 2003; Whitmeyer and Karlstrom, 2007). The NAMG coincides with the final assembly (at ca. 1.6 Ga; Thorkelson et al., 2001; Pourteau et al., 2018) and the following period of relative tectonic stability within the supercontinent Nuna (also called Columbia), and likely reflects Laurentia's central location within the supercontinent (Kirscher et al., 2021). Accordingly, the predominance of NAMG-age detrital zircon grains within western Laurentian strata such as the Mesoproterozoic, ca. 1.47–1.38 Ga Belt (Purcell in Canada) Supergroup (Fig. 2.1A; Ross et al., 1992; Ross and Villeneuve, 2003; Lewis et al., 2010; Box et al., 2020; Brennan et al., 2021), the ca. 1.49–1.44 Ga Hess Canyon Group in southwestern Laurentia (and correlative Trampas basin strata; Doe et al., 2012, 2013; Jones et al., 2015), and the ca. 1.46–1.42 Ga PR1 basin in northwestern Laurentia (Medig et al., 2014), is often interpreted to require non-Laurentian source terranes. Potential source terranes with evidence of NAMG-age magmatism include western Tasmania (Halpin et al., 2014; Mulder et al., 2015), Australia (Reid and Payne, 2017), Antarctica (Goodge et al., 2017) and/or parts of south China (Hainan Island with or without the Cathaysia Block; Li et al., 1995; Xu et al., 2019; Cawood et al., 2020). Consequently, these terranes are commonly located adjacent to western Laurentia within Nuna supercontinent reconstructions.

However, within the Priest River and Clearwater complexes in the northwestern United States of America, the westernmost pre-Belt Supergroup rocks of the Clearwater Block are exposed (Fig. 2.1B), including a small sliver of ca. 1.58 Ga (NAMG-age) granitic gneiss (the Laclede Gneiss; Evans and Fischer, 1986). The ca.

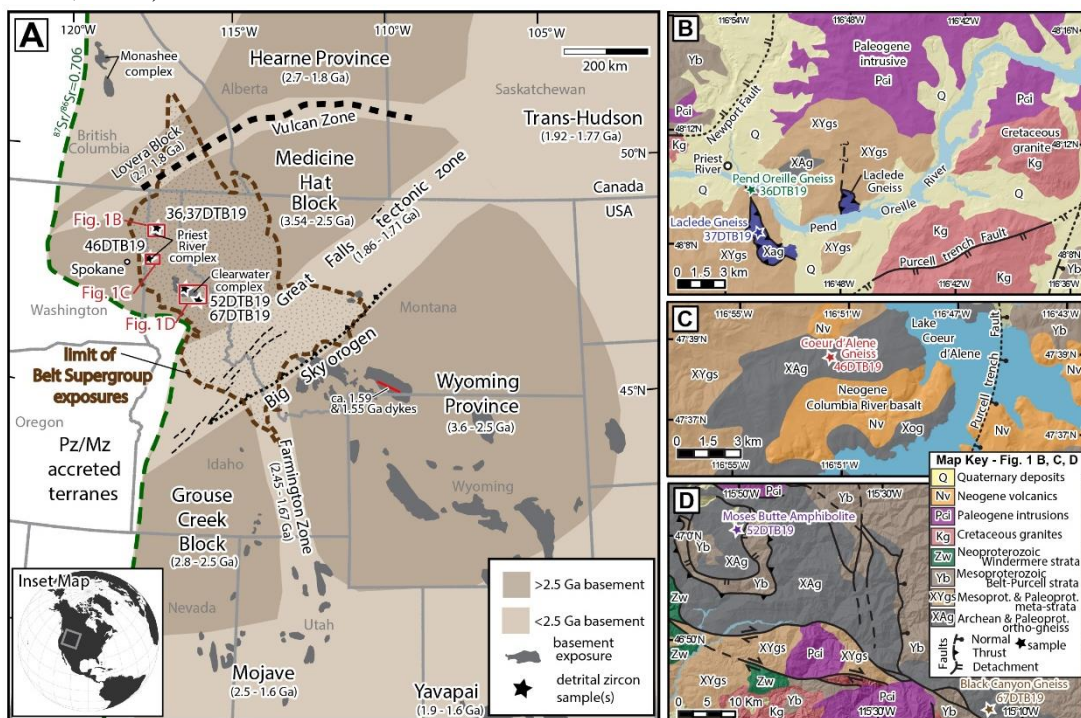
1.58 Ga Laclede Gneiss is the only NAMG-age granitoid in western North America. It has been interpreted as a (i) stranded block of a Proterozoic conjugate continent from Australia, Antarctica, or South China (Ross et al., 1992; Doughty et al., 1998; Goodge et al., 2017; Xu et al., 2019) or (ii) a remnant of a circum-Laurentian orogenesis during the assembly of supercontinent Nuna (Furlanetto et al., 2013). Others (Lewis et al., 2010; Rogers et al., 2018) have also suggested that the Laclede Gneiss might represent a Laurentian source for the ca. 1.7–1.5 Ga detrital zircon grains within parts of the Belt Supergroup, negating the necessity for a western, non-Laurentian source in paleographic reconstructions.

However, apart from U–Pb geochronology studies (Evans and Fischer, 1986; Doughty et al., 1998; Vervoort et al., 2016), Neoproterozoic–Paleoproterozoic Lu–Hf datasets, and some whole-rock geochemistry (Buddington et al., 2016), a multi-isotopic, integrated comparison of the Neoproterozoic–Mesoproterozoic Clearwater Block crystalline basement and overlying Paleoproterozoic–Mesoproterozoic strata does not exist. Consequently, many speculations about the Clearwater Block’s paleogeographic importance remain untested. Here, we present the results of sensitive high-resolution ion microprobe (SHRIMP) U–Pb, secondary ionization mass spectrometry (Cameca SIMS 1280)  $\delta^{18}\text{O}$ , and laser-ablation inductively coupled mass spectrometry (LA-ICP-MS) Lu–Hf and trace element analyses for zircon from ca. 2.67, 1.86 and 1.58 Ga igneous rocks from the Clearwater Block. These new results, along with existing constraints (Evans and Fischer, 1986; Doughty et al., 1998; Lewis et al., 2010; Buddington et al., 2016; Vervoort et al., 2016), provide robust characterization of the westernmost exposed basement beneath the Mesoproterozoic Belt Basin. These rocks include a unique record of NAMG-age magmatism and allow evaluation of competing models for western Laurentia’s assembly and supercontinent Nuna’s configuration/evolution.

## 2.2. Geologic Background

### 2.2.1. The Priest River complex: American or Australian?

Along much of western North America, within the hinterland of the Cordilleran fold and thrust belt, Paleogene orogenic collapse and development of metamorphic core complexes have exhumed rare exposures of deeply buried crystalline basement beneath thick sequences of Mesoproterozoic to Mesozoic strata (e.g. Dickinson, 2004 and references therein). Within the Priest River complex, along the western extent of the Mesoproterozoic Belt Basin in northern Idaho and southern British Columbia (Fig. 2.1), granitic gneisses were among the first Precambrian rocks in the northwestern Cordillera to be dated by U–Pb geochronology (Clark, 1973; Evans and Fisher, 1986; Doughty et al., 1998). In part, these basement rocks drew early attention because of their sharp contrast with the surrounding extensive, siliciclastic and carbonate strata of the ca. 1.47–1.38 Ga Mesoproterozoic Belt (Purcell in Canada) Supergroup that had long been recognized for its economic and tectonic importance (Harrison, 1972; Ross et al., 1992).



**Fig. 2.1:** A) Simplified tectonic map of western Laurentia (after Vervoort et al., 2016 and references within) showing the extent and general age of basement provinces, the limit of Belt Supergroup exposures, location of ca. 1.59 and 1.55 Ga mafic dykes in southern Montana (Tobacco Root Mountains; Rogers et al., 2018) and location of sampling areas. B) Geologic map of the northern extent of basement rock exposure within the Priest River complex where samples of the Pend Oreille Gneiss (36DTB19) and Laclede Gneiss (37DTB19) were collected. C) Geologic map of the southern extent of basement rock exposure within the Priest River complex where a sample of the Coeur d'Alene Gneiss (46DTB19) was collected. D) Geologic map of the southeastern extent of the Clearwater complex where a sample of the Moses Butte Amphibolite (52DTB19) and Black Canyon Gneiss (67DTB19) were collected. Geologic maps B, C, and D adapted from Lewis et al. (2012).

Early geochronological analyses on these basement exposures were characterized by analyses showing considerable U–Pb discordance (Table 1). Discordia regression calculations hinted that the western “Belt basement” within the Priest River complex consisted of Neoproterozoic orthogneiss (ca.  $2650 \pm 21$  Ma) which was unconformably overlain by Belt or possibly pre-Belt strata (Doughty et al., 1998). In addition, the data suggested that the NAMG-age “Laclede Gneiss” ( $1576 \pm 13$  Ma by Evans and Fischer, 1986; ca. 1577 Ma by Doughty et al., 1998) was in fault contact with mylonitized lower Belt Supergroup equivalent strata (Doughty et al., 1998; Lewis et al., 2010; 2020; Fig. 2.1B, 2.1C). Based on the similarity of these ages to existing geochronological constraints for eastern Australia (notably the Gawler Craton), it was suggested that the basement rocks of the Priest River complex could be a remnant of eastern Australia left behind during late Proterozoic rifting (Ross et al., 1992; Doughty et al., 1998; Evans et al., 2000).

Within the Priest River complex, a thin (~300 m thick) quartzite (the Gold Cup Quartzite) unconformably overlies Neoproterozoic and Paleoproterozoic crystalline basement and is overlain by Mesoproterozoic Belt Supergroup strata (Doughty and Chamberlain, 2008; Buddington et al., 2016; Lewis et al., 2020). The Gold Cup Quartzite lacks detrital zircon grains younger than ca. 1.74 Ga, and records a Laurentian provenance (Brennan et al., 2021) suggesting deposition prior to the ca. 1.6 Ga final assembly of supercontinent Nuna (Nordsvan et al., 2018; Pourteau et al., 2018; Kirscher et al., 2019; Volante et al., 2020). Basin-wide studies of the overlying Belt Supergroup suggest that the Belt Basin received sediment from both Laurentian and disputed western non-Laurentian sources (e.g. Ross and Villeneuve, 2003). Within Belt Supergroup strata, a Laurentian signature is characterized by major age-peaks at ca. 2.7, 1.8–1.7, and 1.45 Ga, whereas a possible non-Laurentian signature is characterized as having a broad range of detrital zircon ages from ca. 1.9–1.4 Ga, including significant zircon grains with ages that fall within the ca. 1.61–1.49 Ga North American Magmatic Gap (Ross et al., 1992; Ross and Villeneuve, 2003; Box et al., 2020).

However, others (e.g. Lewis et al., 2010) have questioned if exotic sources are necessary to provide the NAMG-age zircon grains within Belt strata, citing the ca. 1.58 Ga Laclede Gneiss within the Priest River complex as a potential source. Rogers et al. (2018) also advocated that the ca. 1.59 Ga ( $1590 \pm 3$  Ma) Mammoth and ca. 1.55 Ga ( $1551 \pm 5$  Ma) Ramshorn Creek mafic dykes, located nearby in southern Montana,

may have potentially eroded felsic equivalents that could have been additional Laurentian detrital zircon sources. Rogers et al., (2018) correlated the plume-related chemistry of the ca. 1.59 Ga Mammoth dykes to geochemically similar, coeval rocks in the Gawler Craton and Curnamona Province of southeastern Australia, which they consider connected as one single large igneous province (LIP) within Nuna.

### **2.2.2. The Clearwater Block: A Priest River and Clearwater complex connection**

Approximately 150 km south/southeast of the the Priest River complex, pre-Belt Supergroup crystalline basement rocks are also exposed in the Clearwater complex (Fig. 2.1D; Reid et al., 1973; Doughty and Chamberlain, 2007; Vervoort et al., 2016). Early work on Clearwater complex crystalline basement rocks (Table 1) gave discordant geochronology results suggesting ca. 2.1–1.8 Ga crystallization ages (Reid et al., 1973). Follow-up geochronology (Doughty and Chamberlain, 2007) also showed significant Pb-loss, but supported the presence of ca. 1.8 Ga igneous (anorthosite) rocks within the Clearwater complex. Limited geochronology results from an amphibolite exposed within the granitic basement also hinted at the occurrence of ca. 1.59 Ga mafic magmatism based on the two oldest, slightly reverse discordant analyses out of six total analyses, with the younger four analyses suggesting ca. 1.38 Ga metamorphic growth (Doughty and Chamberlain, 2007). Overall, these results reflect the similar timing of pre-Beltian magmatism within the Priest River and Clearwater complexes.

Subsequent extensive geochronological study of the Priest River and Clearwater complexes by Vervoort et al. (2016) further solidified the similarity between the ages of basement rocks exposed within the complexes. Vervoort et al., (2016) dated 18 orthogneiss exposures across the Priest River (n = 2) and Clearwater (n = 16) complexes. All samples yielded ages that fall into two tightly defined age ranges of ca. 2.67 to 2.65, and 1.88 to 1.84 Ga, indicating two main periods of crustal formation beneath the western portion of the Belt Basin. Based on the similarity of these ages, Vervoort et al. (2016) suggested that the Priest River and Clearwater complexes expose crystalline basement of the same crustal block, which they named the Clearwater Block. The relationship between the Clearwater Block and the adjacent Medicine Hat Block to the east is not well known (Vervoort et al., 2016; Gifford et al., 2020).

Paleoproterozoic xenocrystic zircons within Phanerozoic plutonic rocks led Foster et al., (2006) to speculate that a separate ca. 2.4–1.5 Ga “Selway” terrane comprises the southern portion of the Clearwater Block. However, additional studies of the xenocrystic zircon cargo of Phanerozoic plutons (Gaschnig et al., 2013) and exposed basement rocks in the region (Vervoort et al., 2016) did not support this speculation. Consequently, Gaschnig et al. (2013) and Vervoort et al. (2016) refrained from delineating a separate terrane in the southern portion of the Clearwater Block. In addition, geophysical results do not support the interpretation of a separate terrane in the southern portion of the Clearwater Block (e.g. Bedrosian and Feucht, 2014).

### **2.2.3. Adjacent Laurentian basement framework**

Western Laurentia is a complicated mosaic of igneous, metamorphic and sedimentary rocks that range in age from Archean to present and are the result of this region being at or near a continental margin periodically for the past 2.6 billion years (Hoffman, 1988; Whitmeyer and Karlstrom, 2007; Vervoort et al., 2016). East of the Clearwater Block, the Medicine Hat Block contains a mix of Archean (ca. 3.1–2.6 Ga) and Paleoproterozoic (ca. 1.81–1.75 Ga) crust (Gifford et al., 2018, 2020; Ross et al., 1991; Villeneuve et al., 1993). The Paleoproterozoic magmatism within the Medicine Hat block has an evolved Lu–Hf signature indicating reworking of the older Meso- to Neoproterozoic rocks (Gifford et al., 2020).

The Archean Wyoming Craton and Grouse Creek Block are separated from each other by the Farmington Zone and neighbor the Clearwater and Medicine Hat blocks to the south (Whitmeyer and Karlstrom, 2007; Mueller et al., 2011). The Wyoming Craton includes rocks as old as 3.6 Ga and detrital zircon ages as old as 4.0 Ga (Chamberlain et al., 2003; Mueller and Frost, 2006). The Grouse Creek Block in southcentral Idaho and northern Utah may also contain rocks as old as ca. 3.6 Ga preserved only as xenoliths within younger Snake River Plain volcanic rocks (Leeman et al., 1985), but has exposures of ca. 2.67–2.50 Ga felsic and ca. 1.85 Ga mafic magmatic rocks (Strickland et al., 2011; Link et al., 2017). The Farmington Zone contains ca. 2.45 Ga magmatism and ca. 1.67 Ga metamorphism, and likely records the final juxtaposition of the Wyoming Craton, Grouse Creek Block and Mojave Province (Mueller et al., 2011).

Paleoproterozoic (Trans-Hudson-age; ca. 1.9–1.7 Ga) suture/collisional zones separate the main Archean blocks of western Laurentia. The boundary between the Medicine Hat Block and the Wyoming Craton to the south (Mueller et al., 2002) is



delineated by the broad Great Falls Tectonic Zone. The Great Falls Tectonic Zone records major ca. 1.86–1.73 Ga Paleoproterozoic convergent events including the Great Falls Orogeny and the more temporally and spatially restricted Big Sky Orogeny. Generally, Great Falls Tectonic Zone rocks indicate northwest-dipping subduction, closure of an intervening ocean and eventual amalgamation of the Medicine Hat Block to the Wyoming Craton (Gorman et al., 2002; Mueller et al., 2002; Foster et al., 2006; Condit et al., 2015; Gifford et al., 2014; 2018; 2020).

The Clearwater and Medicine Hat blocks are bordered to the north by the subsurface Vulcan structure. The Vulcan structure is likely another Trans-Hudson age feature that may reflect the collision of the southwestern margin of the ca. 2.7–1.8 Ga Hearne Province (the Lovera Block) with the northern margin of the Medicine Hat Block along a north-dipping subduction zone (Ross et al., 1991; Nieuwenhuis et al., 2014). However, differing geophysical interpretations have also resulted in Paleoproterozoic rift (Kanasewich et al., 1969), or south-dipping subduction/collision zone models (Eaton et al., 1999) for the Vulcan structure. North of the Lovera Block, limited geochronologic studies indicate ca. 2.24, 2.03 and 1.84 Ga basement rocks are exposed within the Monashee complex (Crowley et al., 1999).

Following the assembly of the main Archean blocks of western Laurentia during Trans-Hudson-age events, a long-lived accretionary boundary resulted in the addition of the ca. 1.84–1.70 Ga Mojave province, which comprises a mixture of juvenile and evolved components (suggesting the presence of crust as old as ca. 2.5 Ga; Wooden et al., 2012), and the juvenile ca. 1.90–1.70 and 1.65–1.60 Ga Yavapai and Mazatzal provinces along the southern/southwestern margin of Laurentia (Bowring and Karlstrom, 1990; Holland et al., 2020). Collectively, the Mojave-Yavapai-Mazatzal provinces may record progressive accretion of separate, predominantly juvenile arc systems (Whitmeyer and Karlstrom, 2007), or may represent an extensive arc-backarc system that developed along the Australia–Antarctica margin at ca. 1.8 Ga before their accretion to Laurentia at ca. 1.65 Ga (Gibson and Champion, 2019). Mueller et al. (2011) interpreted the ca. 1.67 Ga metamorphism within the Farmington Zone to record collision of Mojave-Yavapai provinces to the southern margin of the Wyoming craton.

## 2.3. Methods

### 2.3.1. Sample preparation

Samples were collected in the context of 1:100,000 scale geologic mapping of the Priest River and Clearwater complexes (Lewis et al., 2007; 2020). Petrographic thin sections were made by Wagner Petrographic, Utah USA. Zircon grains for isotopic analysis were separated from approximately 1 kg individual hand samples using standard techniques consisting of SEL-Frag fragmentation followed by Frantz magnetic and heavy liquid separation at Curtin University, Australia. Unknown zircon grains along with reference zircons OGC, 610, M257, 91500, and Z1 were mounted in epoxy resin within 6 mm of the center of 25 mm diameter disks and were polished to expose grain centers. Prior to analysis, all mounts were imaged in reflected and transmitted light on a Zeiss Axio Imager.M2m optical microscope and cathodoluminescence (CL) imaged using a MIRA3 variable-pressure field-emission scanning electron microscope (SEM) at the Microscopy and Microanalysis Facility at the John de Laeter Centre (JdLC), Curtin University for analysis targeting.

### 2.3.2. A threefold analytical approach: Measurement of U–Pb, $\delta^{18}\text{O}$ , Lu–Hf, and trace element zircon values

U–Pb dating was carried out on the SHRIMP II at Curtin University. A 25  $\mu\text{m}$  diameter spot size was used for all the analysed grains with a primary beam current of 2.5–3.0 nA. Data for each spot were collected in sets of 6 scans through the mass range of  $^{196}\text{Zr}^{2}\text{O}^{+}$ ,  $^{204}\text{Pb}^{+}$ , background,  $^{206}\text{Pb}^{+}$ ,  $^{207}\text{Pb}^{+}$ ,  $^{208}\text{Pb}^{+}$ ,  $^{238}\text{U}^{+}$ ,  $^{248}\text{ThO}^{+}$ ,  $^{254}\text{UO}^{+}$ . The measured isotopic ratios were corrected for common Pb based on the measured  $^{204}\text{Pb}$  and using the common Pb evolution curve of Stacey and Kramers (1975). U–Pb ages were normalized to the M257 zircon standard accepted value of 561.3 Ma (Nasdala et al., 2008). In addition to primary standard M257, secondary reference standard (OGC) analyses were interspersed between unknown analyses during each session (one standard every three to five unknowns). The weighted mean of 18 OGC analyses yielded a date of  $3466.6 \pm 3.1$  (MSWD = 1.6) which is indistinguishable from accepted OGC reference age of  $3465.4 \pm 0.6$  Ma (Stern et al., 2009). All isotopic measurements were reduced, processed and interpreted using SQUID II, and IsoplotR programs (Vermeesch, 2018; Bodorkos et al., 2020; Ludwig, 2003). All uncertainties reported within this text are at  $2\sigma$  and include errors from U–Pb calibration based on the reproducibility of U–Pb standard measurements, counting statistics and the common-

Pb correction. Throughout the text,  $^{207}\text{Pb}/^{206}\text{Pb}$  dates are used for all analyses besides sample 52DTB19 (Moses Butte Amphibolite). For sample 52DTB19, the  $^{206}\text{Pb}/^{238}\text{U}$  and Concordia age values are also considered due to the ambiguity in the best isotopic ratio for dates between ca. 1.5 and 1.0 Ga (Ludwig, 1998), which is discussed further in the results.

**Table 2.1**

Existing and new results of studied Neoproterozoic–Mesoproterozoic magmatism within the Clearwater Block, western Laurentia.

Geologic Unit	Analysis method	Isotopic data	Values reported	Results	Reference
<u>Pend Oreille Gneiss</u>					
PR-93-747B	Dissolution TIMS	U-Pb (n = 3)	discordia upper intercept	2651 ± 21 Ma	Doughty et al., 1998
36DTB19	SIMS-SHRIMP	U-Pb (n = 20)	discordia intercept estimates	ca. 2.7 Ga crystallization, ca. 1.8 Ga metamorphism, or a possible paragneiss origin	this work
<u>Black Canyon Gneiss</u>					
08RAB025	LA-ICPMS	U-Pb (n = 20)	weighted mean of 20 analyses	2663 ± 8 Ma	Vervoort et al., (2016)
67DTB19	SHRIMP	U-Pb (n = 20)	weighted mean of 11 concordant analyses	2665 ± 4 Ma	this work
	SIMS	O- isotopes (n = 20)	weighted mean of 17 robust analyses	$\delta^{18}\text{O} = 5.42 \pm 0.08 \text{‰}$	this work
	LA-SS-ICPMS	Lu-Hf (n = 20)	three most concordant analyses	$\epsilon\text{Hf} = +3.5$ to $-0.5$	this work
	LA-ICPMS	trace element abundances (n = 20)	abundances (ppm)	see Fig. 8A	this work
<u>Coeur d'Alene Gneiss</u>					
09RMG07, 07RMG22	LA-ICPMS	U-Pb (n = 20)	weighted mean of 20 analyses	1862 ± 3, 1867 ± 4 Ma	Vervoort et al., (2016)
67DTB19	SIMS-SHRIMP	U-Pb (n = 20)	weighted mean of 5 concordant analyses	1857 ± 5 Ma	this work
	SIMS	O- isotopes (n = 20)	weighted mean of 17 robust analyses	$\delta^{18}\text{O} = 5.47 \pm 0.07 \text{‰}$	this work
	LA-ICPMS	trace element abundances (n = 20)	abundances (ppm)	reported in supplementary	this work
<u>Laclede Gneiss</u>					
Not assigned	Dissolution TIMS	U-Pb (n = 3)	discordia upper intercept	ca. 1540 Ma	Clark, 1973
Not assigned	Dissolution TIMS	U-Pb (n = 4)	discordia upper intercept	1576 ± 13 Ma	Evans and Fischer, 1986
PR-93-731	Dissolution TIMS	U-Pb (n = 3)	single most concordant analysis	1577.4 ± 2.3 Ma	Doughty et al., 1998
37DTB19	SIMS-SHRIMP	U-Pb (n = 24)	weighted mean of 22 concordant analyses	1581 ± 3 Ma	this work
	SIMS	O- isotopes (n = 20)	weighted mean of 17 robust analyses	$\delta^{18}\text{O} = 5.15 \pm 0.13 \text{‰}$	this work
	LA-SS-ICPMS	Lu-Hf (n = 20)	range of 18 analyses excluding 2 outliers	$\epsilon\text{Hf} = -5.5$ to $-9.7$	this work
	LA-ICPMS	trace element abundances (n = 20)	abundances (ppm)	see Fig. 8B	this work
<u>Moses Butte Amphibolite</u>					
GM-01-2	SIMS-SHRIMP	U-Pb (n = 6)	weighted mean of 2 oldest analyses	1587 ± 9 Ma (crystallization) 1361 ± 13 Ma (metamorphic growth)	Doughty and Chamberlain, 2007
52DTB19	SIMS-SHRIMP	U-Pb (n = 27)	weighted mean of 2 youngest analyses various metrics discussed in text	ca. 1417 to 1350 Ma metamorphic growth	this work
	SIMS	O- isotopes (n = 20)	weighted mean of 17 robust analyses	$\delta^{18}\text{O} = 4.52 \pm 0.31 \text{‰}$	this work
	LA-ICPMS	trace element abundances (n = 20)	abundances (ppm)	reported in supplementary	this work

Abbreviations: TIMS-Thermal Ionization Mass Spectroscopy, LA-Laser ablation, SS-Split stream, ICPMS-Inductively Coupled Plasma Mass spectrometry, SHRIMP-Sensitive High Resolution Ion Micro Probe, SIMS-Secondary Ionization Mass Spectroscopy.

After SHRIMP analysis, the sample mounts were thoroughly cleaned, and briefly repolished with a 3  $\mu\text{m}$  diamond paste, prior to being coated with an approximately 30-nm-thick Au coating. Subsequent zircon oxygen isotopic

compositions were measured using a Cameca IMS 1280 at the Centre for Microscopy, Characterization and Analysis (CMCA) facility at the University of Western Australia following procedures described by Hartnady et al., (2020). Oxygen analysis spots were placed within the same crystal domain as the SHRIMP analysis spots as identified by CL imaging but did not overlap and generally an approximately 10  $\mu\text{m}$  gap or larger was left between analysis spots where feasible. Oxygen results were processed with a CMCA in-house excel macro. Oxygen isotope compositions are reported in the conventional delta notation; expressed as  $\delta^{18}\text{O}$ , reflecting the permil (‰) deviation in the ( $^{18}\text{O}/^{16}\text{O}$ ) isotope ratio of the sample relative to average seawater (Vienna Standard Mean Ocean Water; VSMOW; Baertschi, 1976).

Instrumental drift and fractionation were determined through analysis of reference zircon 91,500 with OGC and M257 as secondary reference standards. During our first of two analytical runs, unknown samples 67DTB19, 46DTB19 and 37DTB19 were analysed with an approximately 20  $\mu\text{m}$  diameter, 2.5nA,  $\text{Cs}^+$  ion-beam. During this run primary reference standard 91,500 ( $n = 14$ ) yielded a  $\delta^{18}\text{O}$  weighted mean of  $9.94 \pm 0.08\text{‰}$  (MSWD = 0.63) which is indistinguishable from the accepted value of  $9.90 \pm 0.30\text{‰}$  (Wiedenbeck et al., 2004) and consequently did not require application of any drift correction. Secondary reference standards yielded  $\delta^{18}\text{O}$  weighted means of  $5.68 \pm 0.21\text{‰}$  (OGC) and  $13.99 \pm 0.06\text{‰}$  (M257) which are also indistinguishable from the accepted values of  $5.88 \pm 0.06\text{‰}$  and  $13.93 \pm 0.22\text{‰}$  (Nasdala et al., 2008; Petersson et al., 2019).  $^{16}\text{O}^1\text{H}/^{16}\text{O}$  values were also monitored to evaluate secondary water update facilitated by radiation damage within the zircon (e.g. Liebmann et al., 2021). Reference standard  $^{16}\text{O}^1\text{H}/^{16}\text{O}$  values ranged from 0.0006 to 0.0010. Only 5/60 analyses of the unknowns exceeded this range with no obvious effect on  $\delta^{18}\text{O}$  values. Nevertheless, these five analyses are disregarded due to their elevated  $^{16}\text{O}^1\text{H}/^{16}\text{O}$  values recording possible alteration of their oxygen isotopic compositions.

Due to smaller zircon size, sample 46DTB19 was analysed during a separate run with a more targeted, approximately 10  $\mu\text{m}$  diameter, 1.0nA,  $\text{Cs}^+$  ion-beam. Slight sample topology observed within the 91,500 reference material (and inability to polish further without removing the unknown grains) rendered it unfavourable as a primary standard (e.g. Kita et al., 2009), and consequently M257 was utilized as the primary reference standard. M257 yielded a  $\delta^{18}\text{O}$  weighted mean ( $n = 6$ ) of  $13.99 \pm 0.12\text{‰}$  (MSWD = 0.62) which is indistinguishable from the accepted value of  $13.93 \pm 0.22\text{‰}$  (Nasdala et al., 2008) and consequently no drift correction was applied. Secondary

reference standard OGC yielded a  $\delta^{18}\text{O}$  weighted mean of  $5.86 \pm 0.18\text{‰}$  which is also indistinguishable from the accepted values of  $5.88 \pm 0.06\text{‰}$  (Petersson et al., 2019). All unknown analyses ( $n = 8$ ) fell within the  $^{16}\text{O}^1\text{H}/^{16}\text{O}$  range given by the standards (0.0009 to 0.0013).

Final analysis consisted of measuring U–Pb, Lu–Hf, and trace element isotopes using laser ablation (single and) split-stream inductively coupled mass spectroscopy (LA-SS-ICP-MS) at the GeoHistory Facility in the John de Laeter Centre at Curtin University, Perth Australia across two analytical sessions. The first analysis session consisted of ablation with 38- $\mu\text{m}$  analysis spot generated by a Resonetics RESolution M50-A excimer laser. From this ablated material, U–Pb isotopes were measured on an Agilent 7700 s quadrupole ICP-MS for age determination, while Lu–Hf isotopes were simultaneously measured on a Nu Instruments NP II MC-ICP-MS. During this session, 38  $\mu\text{m}$  analysis spots were placed over both the SHRIMP and SIMS analysis spots where possible.

Subsequently, during a second laser ablation analytical session (LA-ICP-MS), trace elements abundances were measured on an Agilent 8900 QQQ ICP-MS using a 20  $\mu\text{m}$  spot generated by a Resonetics RESolution M50-A excimer laser which was placed over the same zircon crystal domain as the prior analyses. Following analysis, data reduction was performed using the software Iolite v.3.6 (Paton et al., 2011). MudTank zircon standard was utilized as the primary standard for Lu–Hf results, while GJ-1 was utilized as a primary reference standard for the trace element results.

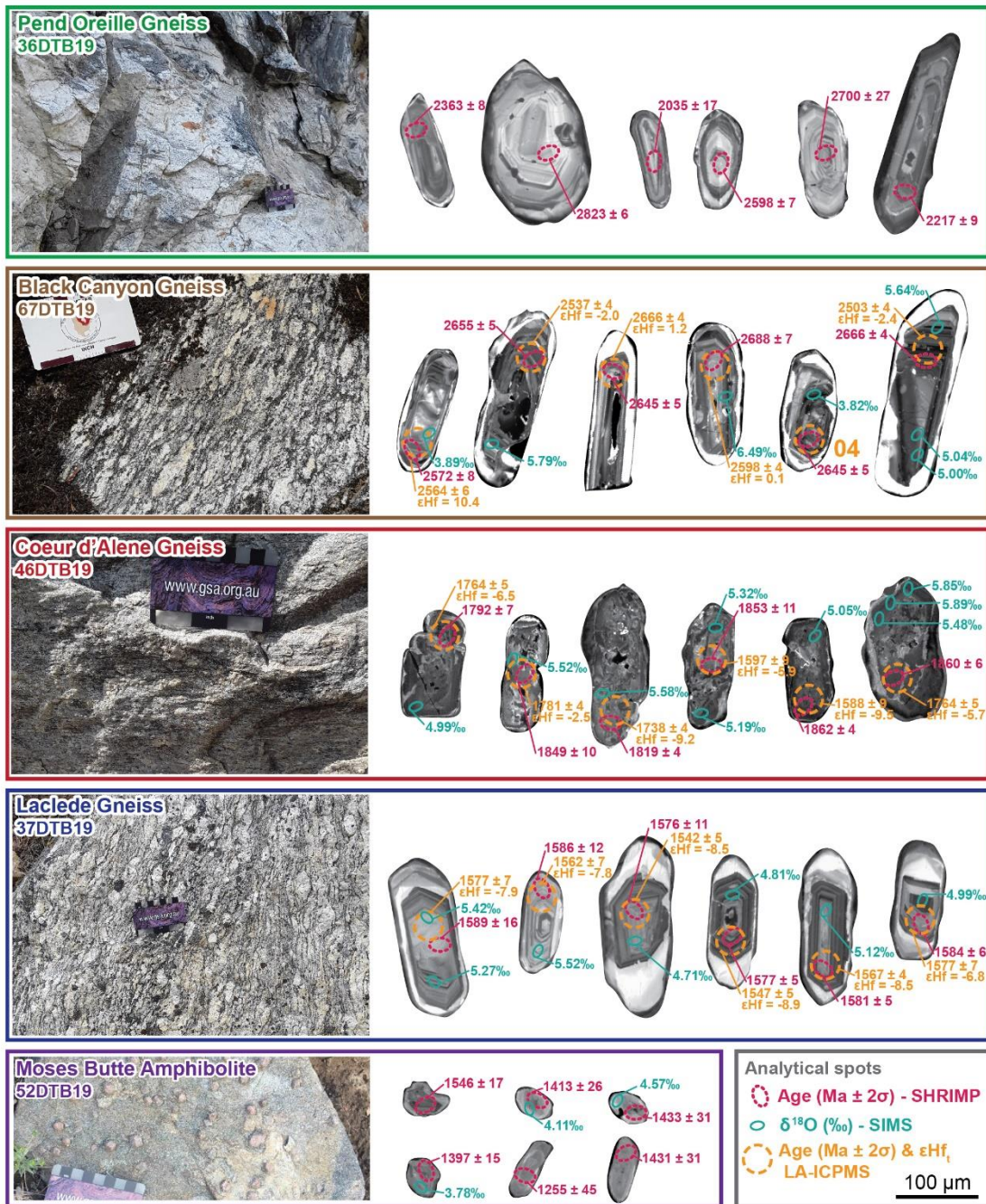
To evaluate the necessity of non-Laurentian sources for strata in the region, new trace element data from detrital zircon grains within overlying Paleoproterozoic Gold Cup Quartzite and Mesoproterozoic lower Belt Supergroup strata are reported. This analytical session also utilized a Resonetics RESolution M50-A excimer laser paired with an Agilent 7700 s quadrupole and Nu Instruments NP II MC-ICP-MS at the GeoHistory Facility in the John de Laeter Centre at Curtin University for simultaneous measurement of U–Pb, Lu–Hf, and trace element abundances. Brennan et al. (2021) report the U–Pb and Lu–Hf results from this detrital zircon analytical session. Additional SHRIMP, SIMS, and LA-ICP-MS analytical details are included in the supplementary information.

## 2.4. Results

### 2.4.1. Sample 36DTB19 – Pend Oreille Gneiss

Within the core of the Priest River complex, the structurally lowest unit exposed is the Pend Oreille Gneiss (Lewis et al., 2020). Doughty et al., (1998) reported U–Pb (thermal ionization mass spectroscopy) data from five chemically abraded zircon grains from this unit that are variably discordant. Their discordia regression suggests a  $2651 \pm 21$  Ma crystallization age, disturbed by an early Paleozoic (ca. 495 Ma) Pb-loss event.

To further evaluate this probable Neoproterozoic basement within the Priest River complex, sample 36DTB19 was collected from a heterogeneous migmatitic biotite gneiss outcrop of the Pend Oreille Gneiss in a similar location (Fig. 2.1B) to Doughty et al. (1998). The zircon grains separated from sample 36DTB19 show concentric zoning in CL imaging consistent with magmatic growth. Thin ( $<5 \mu\text{m}$ ) metamorphic rims are present on some grains but their small size precluded analysis (Fig. 2.2A). Twenty SHRIMP U–Pb analytical results from these grains yielded complex (8% to 90%) discordant results with  $^{207}\text{Pb}/^{206}\text{Pb}$  ages ranging from ca. 2824 Ma to 1634 Ma (Fig. 2.3). The extensive Pb-loss in this sample precluded calculation of a geologically meaningful weighted mean age. A range of Th/U ratios (approximately 0.7 to 0.007) and U (2350 to 350 ppm) abundances were measured (Fig. 2.4). More radiogenic (higher U) grains yielded younger  $^{207}\text{Pb}/^{206}\text{Pb}$  ages. Due to the complexity of this sample, which is discussed further in section 5.1 and 5.2 of the discussion, we also present a more detailed concordia diagram and a thin section microphotograph (Fig. 2.5). The mineralogy observed within this sample (Fig. 2.5B) is relatively simple consisting of abundant lamellar twinned plagioclase, quartz, and biotite.



**Fig. 2.2:** Field photos of the sampled outcrops and cathodoluminescence (CL) images of representative zircon grains from the samples with analytical spots and results indicated. Labeled ages are based on <sup>207</sup>Pb/<sup>206</sup>Pb ratios. Note the scale card in the field photos is approximately 8.5 by 5.4 cm and shows a cm scale bar on the top and inch scale bar on the bottom. Analytical spot key and zircon grain scale bar is in the bottom right.

#### 2.4.2. Sample 67DTB19 – Black Canyon Gneiss

To compare the Archean isotopic composition of the Clearwater Block with the adjacent Medicine Hat Block and Wyoming Craton, the southern and easternmost exposure of Neoproterozoic Clearwater crust with a previously reported crystallization age of  $2663 \pm 8$  Ma (Vervoort et al., 2016) was resampled from along Black Canyon near the southeastern extent of the Clearwater complex (Fig. 2.1D). The Black Canyon



Gneiss, sample 67DTB19, was collected from this Neoproterozoic biotite augen gneiss, and yielded zircon grains with concentric zoning in CL imaging consistent with magmatic growth (Fig. 2.2). Most grains also show bright (low U; e.g. Hancher and Miller, 1993) overgrowths interpreted as metamorphic rims, which likely reflect late Mesozoic or younger metamorphism within the Clearwater complex (e.g. Doughty et al., 2007). Twenty analyses of the oscillatory zoned zircon cores yielded  $^{207}\text{Pb}/^{206}\text{Pb}$  ages (21 to 0% discordant) ranging from ca. 2688 Ma to 2373 Ma (Fig. 2.3), with a relatively limited range of Th/U ratios (0.54 to 0.28; Fig. 2.4). Eleven of these twenty analyses overlap with concordia (within  $2\sigma$  uncertainties) and yield a weighted mean age of  $2665 \pm 4$  Ma (MSWD = 6.6) which is indistinguishable from the age reported by Vervoort et al. (2016). The relatively high MSWD value of the weighted mean age indicates that the calculated weighted mean uncertainty is likely an underestimation of the true imprecision (e.g. Spencer et al., 2016).

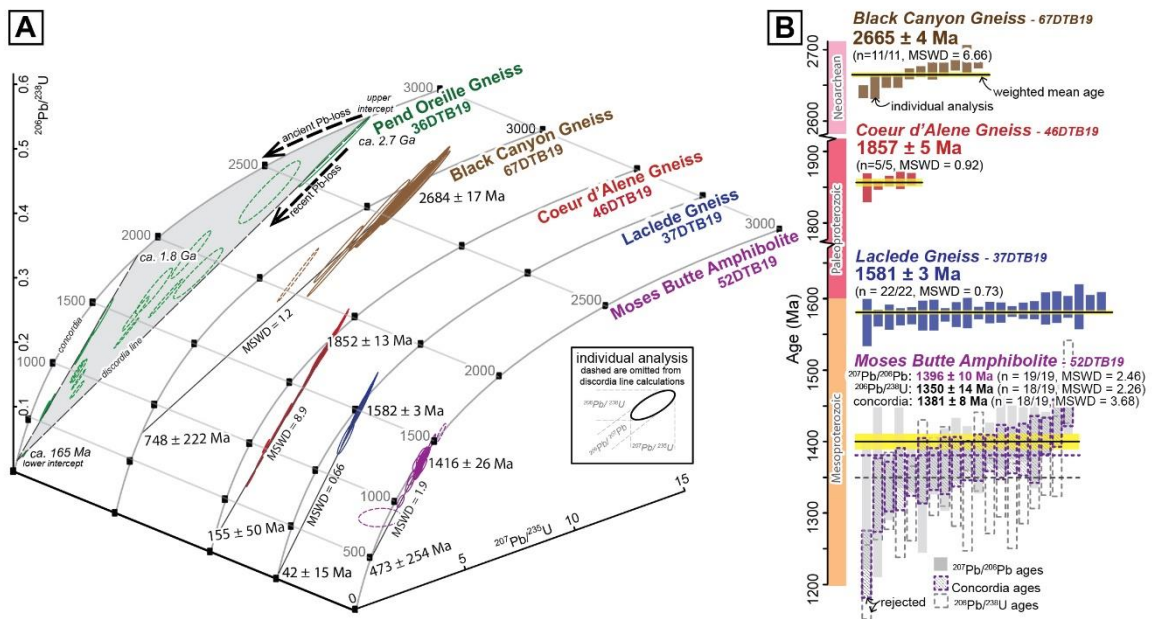
Sixteen robust analyses (2 rejected for elevated  $^{16}\text{O}^1\text{H}/^{16}\text{O}$  values, another 2 rejected as outliers from the weighted mean calculation) of  $\delta^{18}\text{O}$  within the cores of these zircon give a weighted mean of  $5.42 \pm 0.08\text{‰}$  (MSWD = 3.1; Fig. 2.6), which is indistinguishable from accepted  $5.3 \pm 0.6\text{‰}$   $\delta^{18}\text{O}$  value of mantle derived zircons (Valley et al., 1998; Valley, 2003). Eighteen of the twenty Lu–Hf measurements give  $\epsilon\text{Hf}_t$  values from +4.2 to +0.2 when calculated using the preferred 2665 Ma crystallization age, indicating a Lu/Hf composition overlapping and slightly above CHUR (Chondritic Uniform Reservoir; Blichert-Toft et al., 1999; Bouvier et al., 2008). Two additional analyses give outlier  $\epsilon\text{Hf}_t$  values at approximately +6.9 and +13.4 (Fig. 2.7).

#### **2.4.3. Sample 46DTB19 – Coeur d’Alene Gneiss**

To further evaluate the alleged (e.g. Doughty et al., 1998) southeastern Australian affinity of the basement rocks within the Priest River complex, the westernmost exposure of the late Paleoproterozoic Coeur d’Alene Gneiss, within the Priest River complex was sampled along the western shore of Lake Coeur d’Alene (Fig. 2.1C). Vervoort et al. (2016) sampled this same outcrop and report  $1862 \pm 3$  and  $1867 \pm 4$  Ma crystallization ages (LA-ICP-MS). The leucogranite lithology of the Coeur d’Alene Gneiss we sampled yielded zircon grains with poorly preserved and/or absent zoning, rather heterogeneous mottled textures and small localized bright (in CL) patches (Fig. 2.2). Twenty analyses of the most uniform interior regions of these grains yielded  $^{207}\text{Pb}/^{206}\text{Pb}$  ages from 1864 to 1694 Ma, with many of the analyses discordant



and a plot along a Pb-loss line with an  $1852 \pm 13$  Ma upper intercept and a ca. 155 Ma lower intercept (Fig. 2.3A). This lower intercept is similar to the timing of nearby late Jurassic terrane accretion (e.g. Schwartz et al., 2011). Five of the twenty total analyses overlap concordia within  $2\sigma$  uncertainties and yield a weighted mean age of  $1857 \pm 5$  Ma (MSWD = 0.9; Fig. 2.3B), which again is indistinguishable from the ages reported by Vervoort et al. (2016). The zircon grains from sample 46DTB19 yielded the highest U abundances (4472 to 1844 ppm) of all samples, with relatively consistent and low Th/U ratios (0.063 to 0.034; average of concordant grains = 0.051; Fig. 2.4).



**Fig. 2.3:** A) Wetherill concordia diagram of SHRIMP U–Pb results for the sampled units. Dashed analyses are omitted from discordia calculations. B) Weighted Mean calculations for analyses that overlap with concordia. All uncertainties indicated at  $2\sigma$  level.  $^{207}\text{Pb}/^{206}\text{Pb}$  ratios are used for samples 67DTB19, 46DTB19, and 37DTB19. Initial plots and discordia regressions generated with ISOPLOT (Vermeesch, 2018). MSWD – mean standard weighted deviation.

Seventeen robust analyses of  $\delta^{18}\text{O}$  from the same zircon domains as the SHRIMP analyses range from 4.54 to 5.61‰ and give a weighted mean of  $5.47 \pm 0.07\text{‰}$  (MSWD = 3.1; Fig. 2.6), which is indistinguishable from the accepted  $5.3 \pm 0.6\text{‰}$   $\delta^{18}\text{O}$  value for mantle derived zircon (Valley et al., 1998; Valley, 2003). Interestingly, in addition to high U values, the average phosphorous (P), yttrium (Y), and lutetium (Lu) content of these grains (measured during LA-SS-ICP-MS analysis) is over twice that of samples 67DTB19 (Black Canyon Gneiss) and 37DTB19 (Laclede Gneiss). This suggests that the order of magnitude larger ablation volumes during LA-SS-ICP-MS analysis incorporated xenotime (YPO<sub>4</sub>), that was likely present within the zircon grains as either secondary overgrowth along fractures or inclusions. These

contaminant phases were likely avoided during smaller volume U–Pb (SHRIMP) and  $\delta^{18}\text{O}$  (SIMS) analyses but contributed to the spurious (LA-SS-ICPMS) Lu–Hf measurements.

#### **2.4.4. Sample 37DTB19 – Laclede Gneiss**

Within the Priest River complex, a tabular body of megacrystic augen gneiss, the Laclede Gneiss, crops out over  $<9\text{ km}^2$  and lies in fault contact with surrounding metasedimentary strata of the lower Belt Supergroup (Clark, 1973; Doughty et al., 1998; Doughty and Chamberlain, 2008; Lewis et al., 2020). The Laclede augen gneiss is strongly foliated, with layering defined by biotite-rich and -poor zones, and contains conspicuous, subhedral to euhedral orthoclase megacrysts (3–6 cm in length) within a matrix of orthoclase (30–70%), plagioclase (5–35%), quartz (10–15%), biotite (3–7%), along with minor accessory minerals, such as apatite, zircon, epidote, and sphene (Clark, 1967). Prior geochronologic study of the Laclede Gneiss was on exposures north of the Oreille River and indicated disturbance in the U–Pb system resulting from recent Pb-loss (Evans and Fischer, 1986; Doughty et al., 1998). In an effort to minimize the likelihood of similar discordant results, a Laclede Gneiss sample was collected further from the nearby Cretaceous and Paleogene intrusions along the southeastern portion of the outcrop (Fig. 2.1B; Lewis et al., 2020).

This Laclede Gneiss sample yielded zircon grains that in CL imaging showed well-defined oscillatory zoning, with bright (low U) relatively homogenous metamorphic overgrowths (Fig. 2.2). Twenty-three SHRIMP analyses of the most uniform, oscillatory zoned interior regions of these grains yielded  $^{207}\text{Pb}/^{206}\text{Pb}$  ages from 1595 to 1566 Ma (Fig. 2.3), with U values from 2800 to 220 ppm and Th/U values  $>0.24$  consistent with magmatic crystallization (Fig. 2.4). Twenty-two of these analyses overlap concordia within  $2\sigma$  uncertainties and yield a weighted mean age of  $1581 \pm 3\text{ Ma}$  (MSWD = 0.73). This weighted mean value is interpreted as the crystallization age and overlaps with the existing discordia-calculated upper-intercept age-constraint of  $1576 \pm 13\text{ Ma}$  (Fig. 2.3; Evans and Fischer, 1986). A single analysis on the metamorphic rims yielded a ca. 58 Ma ( $^{206}\text{Pb}/^{238}\text{U}$ ) date, which falls between the crystallization age of the nearby  $50 \pm 6\text{ Ma}$  Paleogene pluton (Whitehouse et al., 1992) and ca. 64 Ma peak metamorphism with the Priest River complex (Stevens et al., 2015).

Twenty robust analyses of  $\delta^{18}\text{O}$  from the same zircon domains as the SHRIMP analyses give a weighted mean of  $5.15 \pm 0.13\text{‰}$  (MSWD = 3.4; Fig. 2.6), which is

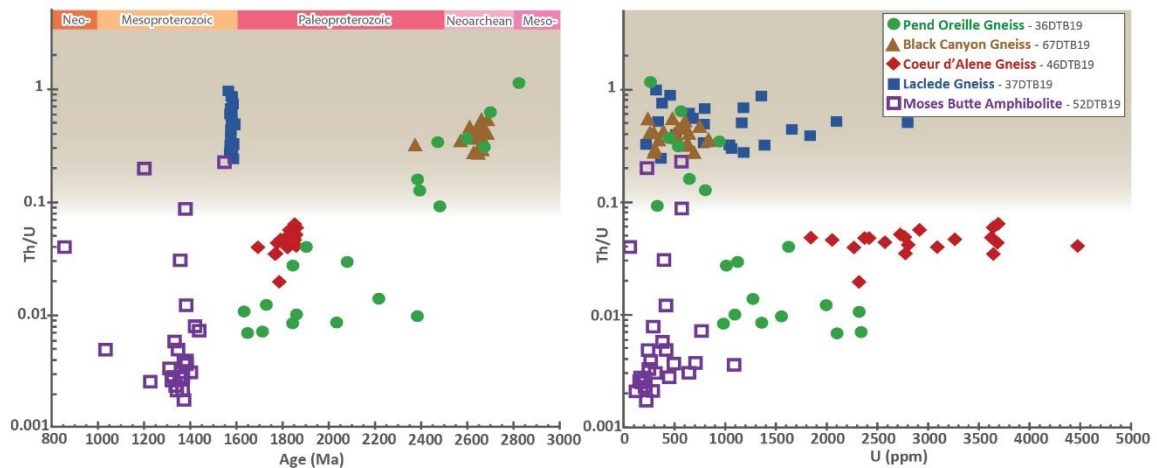
also indistinguishable from accepted  $5.3 \pm 0.6\text{‰}$   $\delta^{18}\text{O}$  value of mantle-derived zircon (Valley et al., 1998; Valley, 2003). Seventeen Lu–Hf measurements give  $\epsilon\text{Hf}_t$  values from -5.4 to -9.3, while three additional analyses give slight outlier  $\epsilon\text{Hf}_t$  values of approximately -5.1, -4.5 and -4.2 (Fig. 2.7).

#### **2.4.5. Sample 52DTB19 – Moses Butte Amphibolite**

Doughty and Chamberlain (2007) reported six zircon analyses from a strongly lineated garnet-hornblende amphibolite collected near Moses Butte within the central portion of the Clearwater complex. Three of these analyses yielded concordant dates of 1520, 1589, and 1584 Ma, which they interpret to record a  $1587 \pm 9$  Ma crystallization age of the mafic protolith. The additional three analyses yielded slightly discordant results suggesting  $1361 \pm 13$  Ma metamorphism. To further evaluate the presence of NAMG-age magmatism within the Clearwater complex, we collected another sample (sample 52DTB19) of the Moses Butte Amphibolite.

Zircon grains recovered from this sample were sparse and small. Most grains showed very faint to absent zoning in CL imaging (Fig. 2.2). Only one recovered grain showed a faint oscillatory zoned core, with a smaller metamorphic overgrowth in CL imaging. Twenty of twenty-six individual SHRIMP U–Pb analyses yielded concordant ages. Nineteen concordant ages on the predominant zircon grains with faint to absent zoning showed Concordia ages from ca. 1440–1229 Ma (Fig. 2.3) with Th/ U ratios from 0.19 to 0.002 (Fig. 2.4). The single oscillatory zoned core yielded a concordant ca. 1546 Ma ( $^{207}\text{Pb}/^{206}\text{Pb}$ ) age with the highest measured Th/U value within this sample of 0.22. The  $^{207}\text{Pb}/^{206}\text{Pb}$  ratios from the nineteen grains with faint to absent zoning yield a  $1396 \pm 10$  Ma (MSWD = 2.46) weighted mean age. Alternative weighted mean calculations based on Concordia ages suggest a  $1381 \pm 8$  Ma (MSWD = 3.68, 1/19 analyses rejected) or  $1350 \pm 14$  Ma (MSWD = 2.26, 1/19 analyses rejected) age using  $^{206}\text{Pb}/^{238}\text{U}$  ratios. The MSWD values of these weighted mean ages suggest underestimated errors, or multiple ca. 1400–1350 Ma populations within in the dataset (e.g. Spencer et al., 2016).

Eight robust analyses of  $\delta^{18}\text{O}$  from the same zircon domains as the SHRIMP analyses range from 5.01 to 4.17‰ and give a weighted mean of  $4.52 \pm 0.31\%$  (MSWD = 23; Fig. 2.6). This is lower than, but slightly overlapping with, the  $5.3 \pm 0.6\%$   $\delta^{18}\text{O}$  value of mantle derived zircon (Valley et al., 1998; Valley, 2003). It is also below the usual 5 to 17‰ value of metabasic metamorphic zircon (Cavosie et al., 2011). The tiny size of these grains precluded robust LA-ICP-MS analysis.



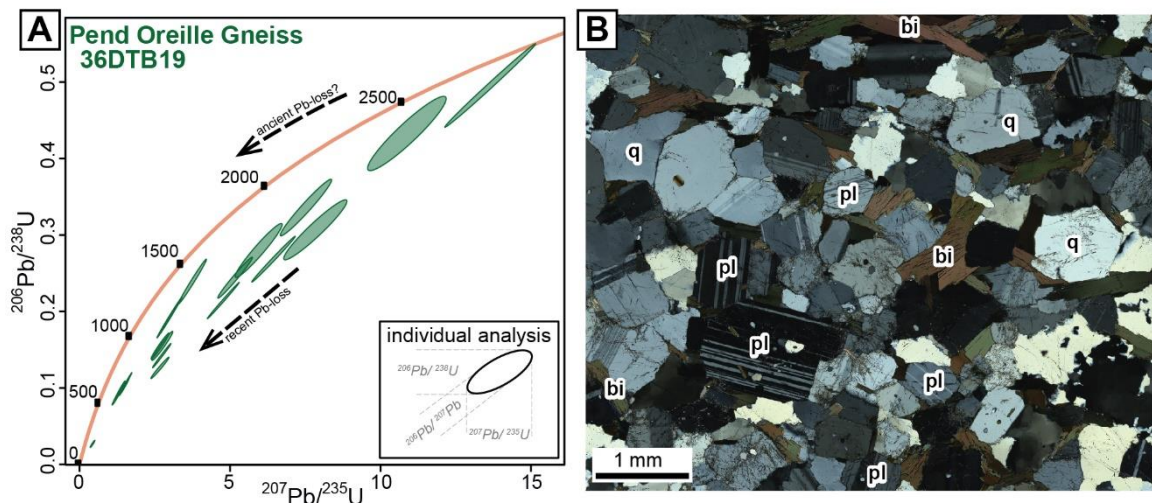
**Fig. 2.4:** Age (Ma) vs. Th/U (left) and U (ppm) vs. Th/U (right) scatterplots for individual SHRIMP U–Pb analyses. Background shading indicates a Th/U ratio of 0.1, which is the lower limit commonly found in igneous zircon (Yakymchuk et al., 2018). Note the vertical (Th/U) axis is logarithmic.

## 2.5. Discussion

### 2.5.1. Neoproterozoic growth of the Clearwater Block

The Neoproterozoic was a time of globally widespread continental growth as evident by the prevalence of ca. 2800–2500 Ma magmatic rocks within many other cratons (e.g. Slave, Yilgarn, Wyoming, etc.; Ivanic et al., 2012; Mole et al., 2019, 2021) as well similar age-peaks within the global detrital zircon record (e.g. Roberts and Spencer, 2015). The proposed mechanisms of crustal generation during the Archean are diverse but are generally placed into two end-member categories. These include modern plate tectonic style plume, rift or subduction processes (e.g. Frieman et al., 2021; Mole et al., 2021), and processes that do not require modern plate tectonics such as delamination, mafic reworking, and so-called “drip tectonics” (e.g. Johnson et al., 2014; Bédard, 2018; Mole et al., 2019). Our study of the Neoproterozoic crust within the Clearwater Block allows us to evaluate several of these potential mechanism for the generation of ca. 2670 Ma crust in this region.

Our sampling (sample 36DTB19) of the Neoproterozoic Pend Oreille Gneiss within the Priest River complex revealed more difficult to interpret U/Pb results than those reported by Doughty et al., (1998). When our results are viewed in Wetherill concordia space, they are consistent with an original ca. 2700 Ma crystallization event, and at least two Pb-loss events, one ancient at ca. 1800 Ma and one more recent in the latest Jurassic to early Cretaceous at ca. 165 Ma (Fig. 2.5A). The relationship between the Th/U ratio and the  $^{207}\text{Pb}/^{206}\text{Pb}$  date of the individual zircon grains within sample 36DTB19, also suggests that the higher U grains, which likely experienced more radiation damage, were preferentially reset during these Pb-loss events (Fig. 2.4). This could indicate that the ca. 495 Ma discordia lower intercept (an age not commonly associated with metamorphism in this region) calculated by Doughty et al., (1998) for the Pend Oreille Gneiss is actually a mixing age between Paleoproterozoic and Cretaceous Pb-loss. However, given the complexity of the sampled gneiss outcrop (Fig. 2.2), it is also possible that our sample 36DTB19 represents a high-grade paragneiss phase containing ca. 2700 and 1800 Ma detrital zircon populations that experienced latest Jurassic to early Cretaceous Pb-loss. The lamellar twinned plagioclase feldspar, quartz, and biotite mineralogy observed within this sample is generally permissible with either a para- or orthogneiss interpretation (Fig. 2.5B).

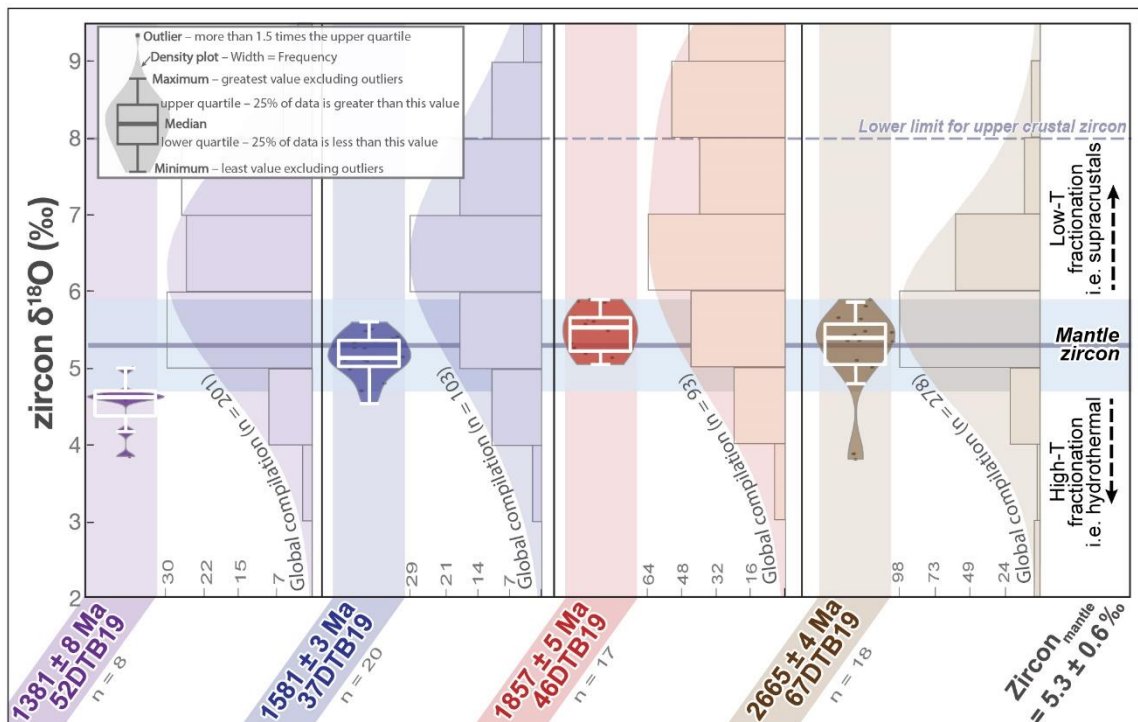


**Fig. 2.5:** A) Wetherill concordia diagram of SHRIMP U–Pb results for sample 36DTB19 indicating the complexity observed. B) A thin-section microphotograph of sample 36DTB19 in cross polarized light showing the main mineral modes of plagioclase (pl), quartz (q), and biotite (bi).

Within the ca. 2670 Ma Black Canyon Gneiss (sample 67DTB19), the augen texture (Fig. 2.2) and analytical results (Figs. 3, 4) are all consistent with a *metaigneous* (orthogneiss) origin. The measured zircon  $\delta^{18}\text{O}$  value of  $5.42 \pm 0.08\%$  from this sample indicates that the southwestern-most region of Archean crust within the

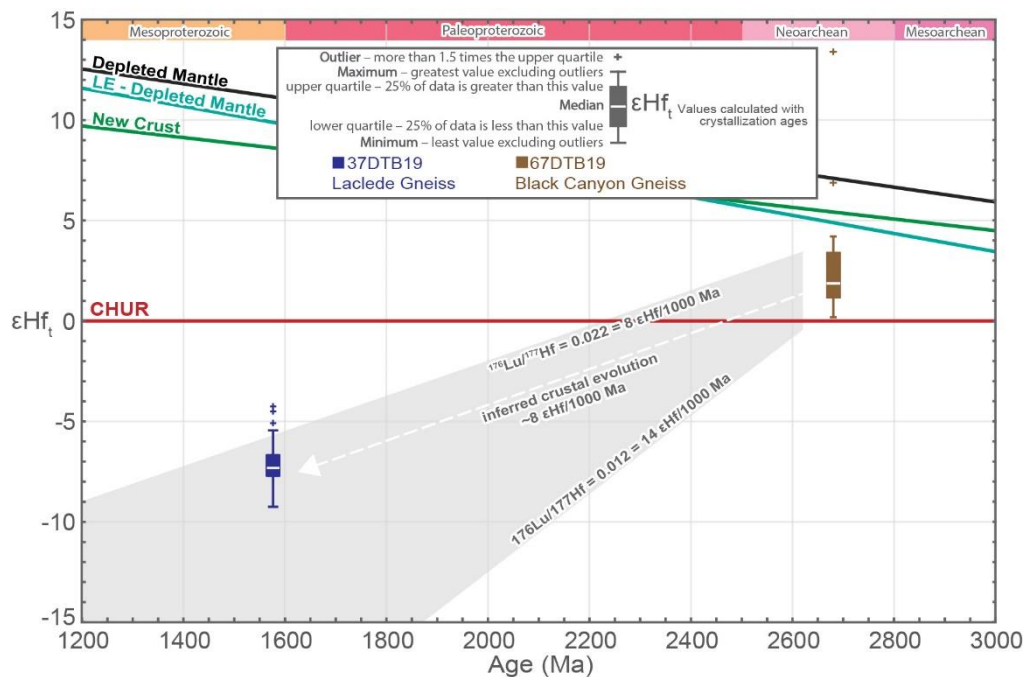


Clearwater Block did not undergo fractionation of O- isotopes since mantle extraction (Fig. 2.6). With some exceptions (e.g. Smithies et al., 2021), most Archean crust gives similar mantle-like  $\delta^{18}\text{O}$  values (Fig. 2.6; Spencer et al., 2014), indicating relatively limited intracrustal recycling of surface materials into magma by melting or contamination (Valley et al., 1998; 2003). The  $\epsilon_{\text{Hf}_t}$  values of these same zircon grains range from +4.2 to +0.2, indicating an overall slightly supra-chondritic (above CHUR; Bouvier et al., 2008) Lu–Hf composition (Fig. 2.7). These  $\epsilon_{\text{Hf}_t}$  values are approximately 3  $\epsilon_{\text{Hf}}$  units below the value of the traditionally applied depleted mantle (DM) evolution line at ca. 2670 Ma. However, this traditional depleted mantle evolution line requires early (ca. 4560 Ma) widespread depletion of the mantle (e.g. Vervoort et al., 1996). If a later timing of widespread mantle depletion, such as widespread mantle extraction at ca. 3800 Ma (e.g. Vervoort and Kemp, 2016; Mole et al., 2021; Mulder et al., 2021) or the proposed new crust evolution curve (Dhuime et al., 2011) is accepted, then the  $\epsilon_{\text{Hf}_t}$  values within the Black Canyon Gneiss are much closer (<2  $\epsilon_{\text{Hf}}$  units) from overlapping the expected value of the mantle reservoir at ca. 2670 Ma (Fig. 2.7).



**Fig. 2.6:** SIMS  $\delta^{18}\text{O}$  results from the sampled units shown in violin plots. A violin plot includes both a box plot which indicates the min, max, median and quartiles of the dataset, and a mirrored density plot that shows the spread of the data. See explanation in the top left. On the vertical axis next to each sample's violin plot is a kernel density estimate plot of  $\delta^{18}\text{O}$  zircon values extracted from the global compilation of Spencer et al. (2014). The global compilation was filtered to include zircon with crystallization ages within 50 million years of each individual Clearwater Block sample, and is plotted with 1‰ bin and band widths. Range of mantle zircon  $\delta^{18}\text{O}$  values from Valley et al. (2008).

Consequently, our Lu–Hf data alone indicates that genesis of the Black Canyon Gneiss could have been derived directly from an only slightly depleted mantle reservoir at ca. 2670 Ma, or via reworking of older crust that was derived from a more depleted mantle reservoir during the Mesoarchean (ca. 3200–2800 Ma). The mantle-like  $\delta^{18}\text{O}$  values for the Black Canyon Gneiss (sample 67DTB19) indicate an absence of O- fractionation from mantle values by either tectonic or hydrothermal processes. The low ( $<0.3$ )  $\text{Eu}/\text{Eu}^*$  [Chondrite normalized  $\text{Eu}/(\text{Sm} \cdot \text{Gd})^{0.5}$ ] ratios from the Black Canyon Gneiss zircon grains (Fig. 2.8) also indicate considerable plagioclase fractionation, and minimal garnet fractionation within the melt during zircon crystallization. This is inconsistent with crystallization within thickened/hydrated continental crust, and suggests crystallization within shallower/drier crust (e.g. Lu et al., 2016; Tang et al., 2021). Based on these constraints alone, generation of the oldest crust within the Clearwater Block does not require crustal reworking of older Mesoarchean crust within a modern plate-tectonic style subduction setting. Consequently, we suggest that the Clearwater Block represents juvenile, mantle-derived, Neoproterozoic crustal growth.



**Fig. 2.7:**  $\epsilon\text{Hf}_t$  values vs.  $(^{207}\text{Pb}/^{206}\text{Pb})$  Age plot for the Black Canyon Gneiss (67DTB19) and Laclede Gneiss (37DTB19). The colored box plots indicate the range of  $\epsilon\text{Hf}_t$  values calculated for each gneiss based on the measured Lu/Hf, and the interpreted crystallization age of the sample. Here and in subsequent  $\epsilon\text{Hf}_t$  vs. Age diagrams, we include three mantle evolution models including the classic Depleted Mantle (Bennett et al., 1993) assuming early mantle depletion at ca. 4560 Ma, the late extraction (LE) depleted mantle model (Vervoort et al., 2016; Fisher and Vervoort, 2018) assuming mantle depletion at ca. 3800 Ma, and the new crust model (Dhuime et al., 2011) which assumes Lu/Hf values of new juvenile crust is best represented by the average of oceanic arc values. CHUR—Chondritic Uniform Reservoir (Blichert-Toft et al., 1999; Bouvier et al., 2008).

However, the adjacent Medicine Hat Block contains both Mesoarchean rocks and locally preserves a sub-chondritic, Lu–Hf, and Sm–Nd, record of Neoproterozoic crustal growth reworking/recycling older Mesoarchean crust (Gifford et al., 2014; 2018; 2020). The thick deposits of Belt Supergroup strata between the Clearwater Block and the Medicine Hat Block prevent the detection of any clear geophysical or geological signature of either a suture/collisional zone or an Archean crustal continuity between the Clearwater and the Medicine Hat blocks (Bedrosian and Feucht, 2014; Gu et al., 2018). Consequently, it would be possible that the Clearwater Block was not associated with the Medicine Hat Block prior to the late Paleoproterozoic assembly of Laurentia via modern plate-tectonic subduction/collisional processes (Whitmeyer and Karlstrom, 2007; Weller and St-Onge, 2017; Wan et al., 2020). After ca. 1800 Ma, the continuity and similar provenance of strata across the Medicine Hat and Clearwater blocks indicate that they had achieved their near-current relative configuration (McMechan et al., 2021; Brennan et al., 2021). Consequently, we only advocate for the usage of a unified Medicine Hat-Clearwater Block (cf. Vervoort et al., 2016; Gifford et al., 2020) terminology for time after ca. 1800 Ma.

### **2.5.2. Paleoproterozoic Tectonism within the Priest River complex**

As previously discussed, the Pend Oreille Gneiss (sample 36DTB19) within the Priest River complex records a composite ca. 2700, 1800, and 165 Ma tectonic history (Fig. 2.5). Within the nearby Coeur d'Alene Gneiss, zircons with low Th/U ratios (0.063 to 0.034), high U abundances (approximately 1800 to 4500 ppm) and rather heterogeneous mottled zircon textures (sample 46DTB19) yield a weighted mean age of  $1857 \pm 5$  Ma and also record a similar timing of recent Pb-loss (ca. 155 Ma). Globally, high  $\delta^{18}\text{O}$  zircon values are common in ca. 1900–1800 Ma rocks (Fig. 2.6) due to widespread crustal reworking associated with Trans-Hudson age orogens (e.g. Spencer et al., 2014). However, the relatively robust  $\delta^{18}\text{O}$  isotopic results from the ca. 1860 Ma Coeur d'Alene Gneiss (sample 46DTB19) record similar mantle-like  $\delta^{18}\text{O}$  values to the ca. 2670 Ma Black Canyon Gneiss (sample 67DTB19). Consequently, we interpret the  $1857 \pm 5$  Ma age for the Coeur d'Alene Gneiss (sample 46DTB19) to record the complete resetting of the U–Pb system of the existing ca. 2670 Ma crust within the Clearwater Block (Vervoort et al., 2016).

The proximity of the Priest River complex to the Vulcan structure suggests that the ca. 1860 Ma event recorded within the Coeur d'Alene Gneiss could be related to the collision of the Medicine Hat-Clearwater Block with the Lovera Block (the



southwestern region of the Hearne Province) along the enigmatic Vulcan structure (Eaton et al., 1999; Nieuwenhuis et al., 2014). Tectonic models for the Vulcan structure include a Paleoproterozoic rift (Kanasewich et al., 1969), and a north- (Ross et al., 1991; Nieuwenhuis et al., 2014) or south-dipping (Eaton et al., 1999) subduction/collision zone. However, mantle-like  $\delta^{18}\text{O}$  isotopic values and the general absence of juvenile magmatism associated with this Paleoproterozoic tectonism within the Priest River complex or in the northern portion of the Medicine Hat Block (Gifford et al., 2020) are most consistent with the model of a north-dipping (away from the Medicine Hat-Clearwater Block) collisional zone within the Vulcan structure.

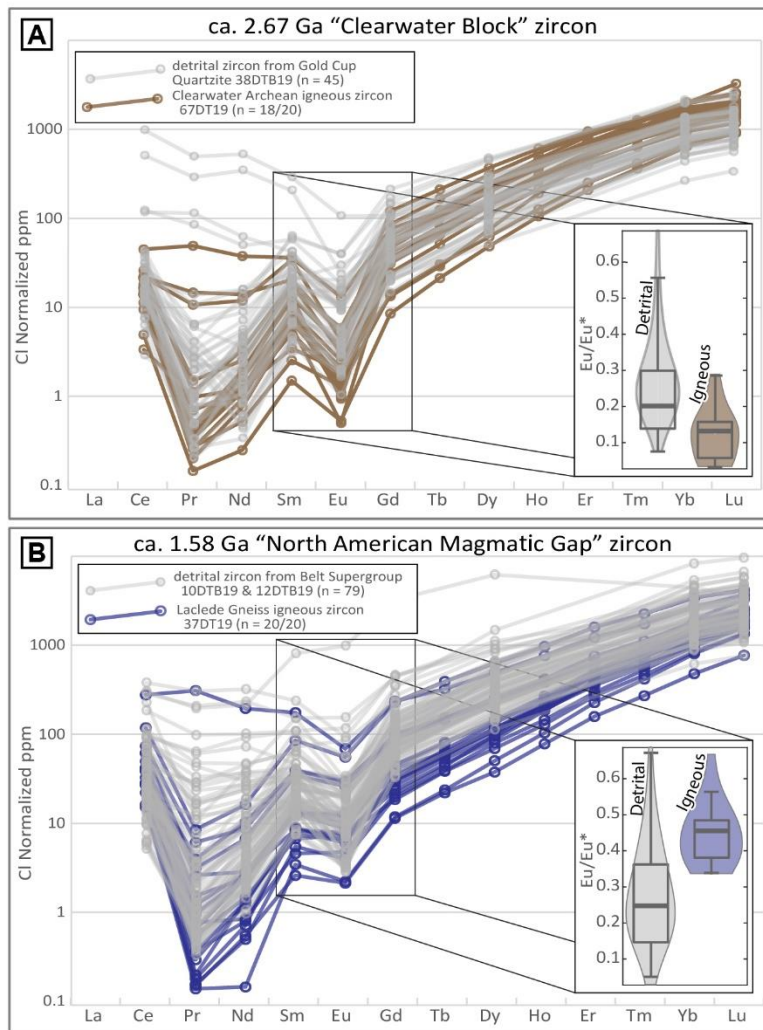
If our speculation that the ca. 1860 Ma tectonism within the Coeur d'Alene Gneiss reflects a collision (and perhaps collisions induced melting) along the Vulcan structure, it occurred coevally with magmatism within the ca. 1900–1700 Ma Great Falls Tectonic Zone (Mueller et al., 2002) but prior to the 1830–1800 Ma terminal phase of the Trans- Hudson orogeny (Corrigan et al., 2009). This is similar to the relative order of events indicated by geophysical cross cutting relationships (Nieuwenhuis et al., 2014). However, more extensive work is needed to evaluate the tectonic relationship of the ca. 1860–1830 Ma metamorphism and magmatism within the Priest River and Clearwater complexes (Buddington et al., 2016; Vervoort et al., 2016), with nearby similar late Paleoproterozoic events recorded in the Vulcan Structure, Great Falls Tectonic Zone, and Monashee Complex (e.g. Ross et al., 1991; Crowley, 1999; Mueller et al., 2002).

### **2.5.3. NAMG-age magmatism within the Clearwater Block**

The results from sample 37DTB19 refine the crystallization age for the Laclede Gneiss from  $1576 \pm 13$  Ma (Evans and Fischer, 1986) to  $1581 \pm 3$  Ma. The Laclede Gneiss clearly falls within the ca. 1610 to 1490 Ma North American Magmatic Gap (Fig. 2.2; NAMG; Ross and Villeneuve, 2003). However, it does not overlap in age with the  $1590 \pm 5$  and  $1551 \pm 5$  Ma mafic dykes in southern Montana (Rogers et al., 2018). Of regional interest, it also confirms that the larger, augen gneiss outcrop south of the Pend Oreille River is correlative to the smaller dated bodies (Fig. 2.1B; Evans and Fischer, 1986; Doughty et al., 1998) on the northern side of the river as mapped (Lewis et al., 2020). The well-preserved oscillatory zoning, Th/U ratios above 0.2, absence of a U (ppm) vs. age trend (Fig. 2.4), and overall mineralogy including large K-spar augen (Fig. 2.2) indicates an igneous/magmatic origin for sample 37DTB19.

The new Lu–Hf and  $\delta^{18}\text{O}$  results also yield additional insights into the origin of this unique Laurentian magmatic event.

The  $\epsilon_{\text{Hf}_t}$  values of the Laclede Gneiss range from -5.1 to -9.3, which indicates an isotopically evolved signature consistent with recycling of Neoproterozoic or Paleoproterozoic crust. Although this sample records evolved Lu–Hf values, the  $\delta^{18}\text{O}$  results are consistent with mantle-like values signifying limited intracrustal recycling of surface materials into the magma by melting or contamination. These Lu–Hf and  $\delta^{18}\text{O}$  results indicate that ca. 1580 Ma magmatism within the Clearwater Block was unlikely subduction or arc-related as previously speculated (e.g. Furlanetto et al. 2013). The zircon trace element results from the Laclede Gneiss also show higher ( $>0.3$ ) Eu/Eu\* ratios than the nearby Neoproterozoic basement (Fig. 2.8). This higher Eu/Eu\* ratio suggests limited plagioclase and significant garnet fractionation within the melt during zircon crystallization, and thus is consistent with derivation from thickened continental crust (e.g. Lu et al., 2016; Tang et al., 2021).



**Fig. 2.8:** A) Chondritic normalized trace element results for the ca. 2670 Ma (sample 67DTB19; shown in brown) igneous rocks of the Clearwater Block, and the same age (overlapping within  $2\sigma$  error) detrital zircon grains from the overlying Gold Cup Quartzite (grey). B) Chondritic normalized trace element results for the ca. 1580 Ma (sample 37DTB19, shown in blue) igneous rocks of the Clearwater Block, and the same age (overlapping within  $2\sigma$  error) detrital zircon grains from the overlying lower Belt Supergroup strata (grey). Inset charts highlight the Eu anomaly calculated as  $[\text{Chondrite normalized Eu}/(\text{Sm} \cdot \text{Gd})^{0.5}]$ .

#### **2.5.4. Crustal Evolution of the Clearwater Block**

A Lu–Hf ‘reworking array’ (e.g., Payne et al., 2016) fit through the 2670 and 1580 Ma Clearwater complex samples intersects with the main ca. 1860–1800 Ma detrital zircon population within the overlying Gold Cup Quartzite (Brennan et al., 2021; Figs. 2.7, 2.9). We interpret these Lu/Hf results to be consistent with a continuous crustal evolution trend recording limited input from a depleted mantle source, at least in the Priest River region, during the ca. 1860 and 1580 Ma events. The slope of this line yields an approximate  $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$  evolution ratio that is higher than the generally suggested value for bulk felsic continental crust ( $^{176}\text{Lu}/^{177}\text{Hf} = 0.018$  to  $0.012$ ) and results in a shallower  $\epsilon\text{Hf}$  vs. time evolution. As mentioned, the available  $\delta^{18}\text{O}$  results from the igneous samples used to define this array are consistent with limited fractionation of O- isotopes from expected mantle values, indicating limited magmatic interaction with low temperature meteoric or high temperature hydrothermal water (Valley et al., 1998), suggesting that a subduction origin for these rocks is not required.

Consequently, the  $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$  ratio suggested by the fit of a reworking array of both the igneous basement and proximally derived overlying detrital zircon samples is plausible given that lower continental (more mafic) crust generally yields higher  $^{176}\text{Lu}/^{177}\text{Hf}$  values ( $0.025$  to  $0.017$ ) than the often assumed values for the bulk or more-felsic upper continental crust (Payne et al., 2016; Spencer et al., 2020). Comparable  $^{176}\text{Lu}/^{177}\text{Hf}$  ratios (e.g.  $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$ , Amelin et al., 1999;  $^{176}\text{Lu}/^{177}\text{Hf} = 0.024$ , Næraa et al., 2012;  $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$ – $0.020$ , Mole et al., 2019) have previously been calculated based on reworking arrays of other Archean provinces. This crustal evolution array suggests that the oldest, ca. 2670 Ma magmatism in the Clearwater Block represents the main crustal growth event within the block, and the subsequent events at ca. 1860 Ma and 1580 Ma likely reworked this Archean lower crustal reservoir, perhaps without the involvement of arc-magmatism. Given the plausibility of this crustal evolution model, the isotopic evidence does not require an Australian origin for the ca. 1580 Ma Laclede Gneiss as speculated by previous studies (e.g. Ross et al., 1992; Doughty et al., 1998).

#### **2.5.5. Ca. 1.4 Ga metamorphism within the Clearwater complex**

The results from the Moses Butte Amphibolite (sample 37DTB19) show no clear evidence for NAMG-age magmatism as most grains yielded textures, Th/U, and  $\delta^{18}\text{O}$  results consistent with metamorphic zircon growth in a hydrothermally affected

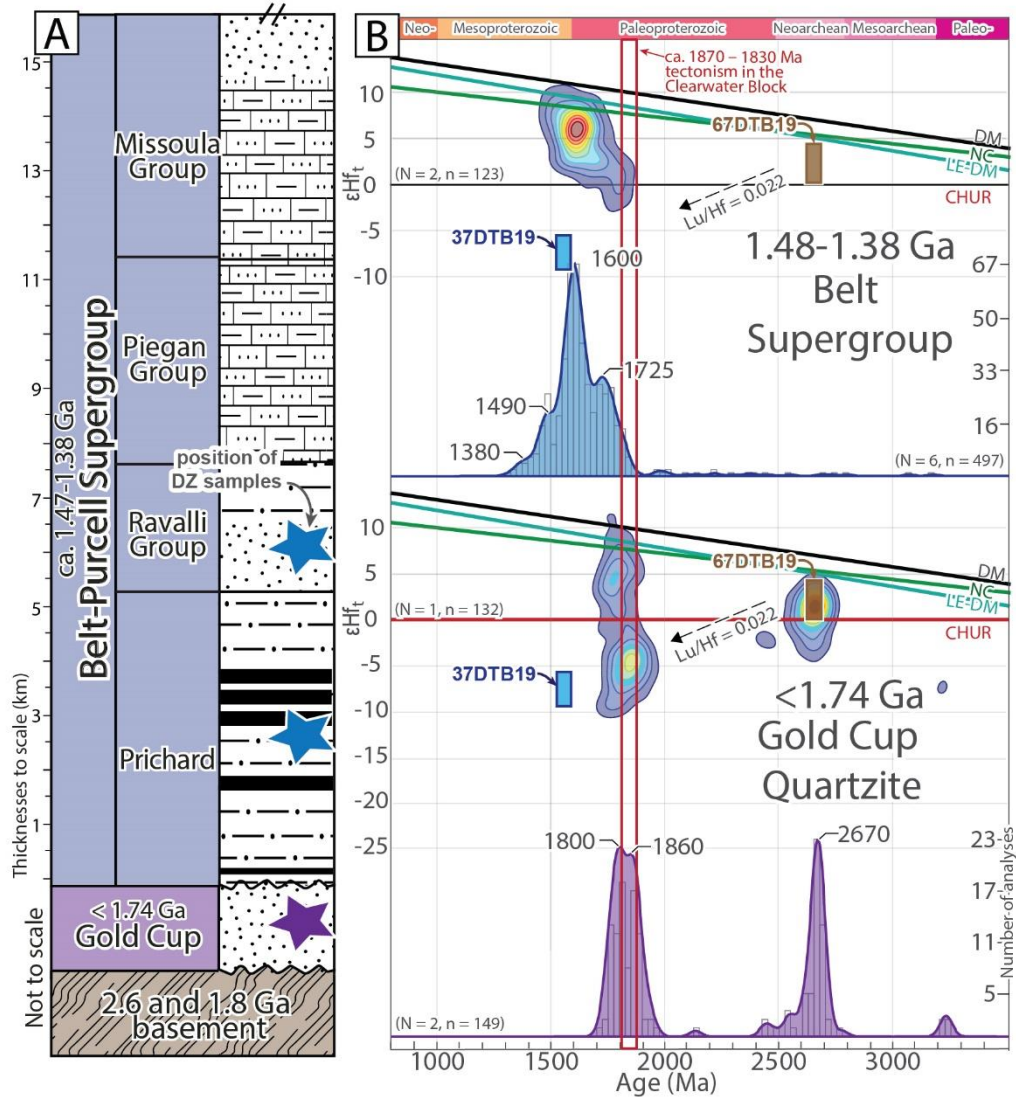
environment (Figs. 2.3, 2.4, 2.6). A single grain showed an oscillatory zoned core (Fig. 2.2), which had the highest Th/U value of 0.22 and yielded a concordant ( $^{207}\text{Pb}/^{206}\text{Pb}$ ) date of  $1546 \pm 17$  Ma. This single grains supports speculations (Doughty and Chamberlain 2007), that the amphibolite's protolith may fall within the NAMG age interval and could be related to the nearby  $1551 \pm 5$  Ma mafic dykes (Rogers et al., 2018). However, the remainder of the dates are interpreted to record metamorphic/hydrothermal zircon growth. The metamorphic event(s) recorded by these grains occurred between ca. 1416 and 1350 Ma, as determined by the oldest discordia upper-intercept age, and the youngest ( $^{206}\text{Pb}/^{238}\text{U}$ ) weighted mean age. Magmatism (and associated metamorphism) of this age is well documented within the Belt Basin (Evans et al., 2001; Doughty and Chamberlain, 2008; Zirakparvar et al., 2010; Nesheim et al., 2012; McFarlane, 2015), which likely provided heat for the hydrothermal zircon growth documented here. This period of magmatism is correlated with rifting near the end of Belt Basin subsidence and is suggested to correspond to global events linked to the initial breakup of supercontinent Nuna/ Columbia (Ross and Villeneuve, 2003; Box et al., 2020; Brennan et al., 2021).

#### **2.5.6. Constraints on detrital zircon provenance**

In addition to providing insights into the genesis of Clearwater Block basement, our results allow evaluation of proposed detrital zircon provenance patterns for the overlying Paleo–Mesoproterozoic strata. In particular, within the Priest River complex, younger than ca. 1800 Ma coarse strata of the Gold Cup Quartzite overlie Clearwater Block Basement (Lewis et al., 2020). Detrital zircon U–Pb and Lu–Hf components from the Gold Cup Quartzite consist of two main age components. An older ca. 2670 Ma population with CHUR-like  $\epsilon\text{Hf}_t$  values concentrated around +1, and younger ca. 1900–1750 Ma grains, with main age-peaks at ca. 1860 and 1800 Ma. These Paleoproterozoic zircon grains have a wide range of  $\epsilon\text{Hf}_t$  values with the greatest density showing evolved  $\epsilon\text{Hf}_t$  values between -4 and -8 (Fig. 2.9; Brennan et al., 2021). These isotopic results from the detrital zircon grains, including the trace element constraints for sample 67DTB19, are a good match for the basement rocks within the Clearwater Block and are consistent with our suggested crustal evolution trend. These similarities support prior interpretations that the Gold Cup Quartzite was sourced from proximal Laurentian basement sources primarily within the Clearwater and Medicine Hat Blocks (Doughty and Chamberlain, 2008; Buddington et al., 2016; Brennan et al., 2021).

In the Priest River region, finer-grained strata of the ca. 1480–1380 Ma Belt Supergroup overlie strata of the Gold Cup Quartzite. Brennan et al. (2021) sampled Prichard and Ravalli Group strata from the stratigraphically lower portion of the Belt Supergroup approximately 50 km to the west of the Priest River complex basement exposures. These samples contained a spread of ca. 1750–1470 Ma detrital zircon grains, including a significant portion of grains with ages within the ca. 1610–1490 Ma NAMG. These NAMG-age grains have predominately juvenile ( $+\epsilon\text{Hf}_t$ ) values. Consequently, the evolved  $\epsilon\text{Hf}_t = -5.1$  to  $-9.3$  values within the ca. 1580 Ma Laclede Gneiss, make it a poor Lu–Hf match for the NAMG-age detrital grains within nearby Belt Supergroup strata (Fig. 2.9). In addition to the Lu–Hf mismatch, the trace element data from the Laclede Gneiss (sample 37DTB19) also match poorly with that of the detrital grains (Figs. 2.8, 2.9). Consequently, interpretations of an exotic source for these detrital grains (Ross and Villeneuve, 2003; Fanning et al., 2009; Box et al., 2020; Brennan et al., 2021) remains a valid hypothesis. Additionally, the non-Laurentian source for the similar, juvenile ( $+\epsilon\text{Hf}_t$ ) NAMG-age detrital zircon populations found in coeval ca. 1450 Ma Hess Canyon, Trampas basins in southwestern Laurentia also remains valid (Doe et al., 2012, 2013; Jones et al., 2015).

Southeastern Australia, notably the Gawler Craton, records widespread magmatism in pulses from ca. 1900 to 1500 Ma, which include: 1) the ca. 1850–1790 Ma Donington–Myola Suite (Fanning et al., 1988; Reid et al., 2008), 2) the ca. 1770–1740 Ma McGregor Volcanics and associated suites (Fanning et al., 1988), 3) the 1620–1605 Ma St Peter Suite (Swain et al., 2008), 4) the 1600–1575 Ma Gawler Range Volcanics–Hiltaba suite (Fanning et al., 2009), and 5) granites as young as ca. 1500 Ma in the Spilsby Suite (Fanning et al., 2009). The ca. 1620–1605 Ma St Peter Suite and ca. 1600–1575 Ma Hiltaba Suite and Gawler Range Volcanics show juvenile ( $+\epsilon\text{Hf}_t$ ) compositions (Swain et al., 2008; Reid and Payne, 2017) and are consequently a much better source match for the detrital zircon grains within Belt Supergroup strata than the Laclede Gneiss. Thermochronology ( $^{40}\text{Ar}/^{39}\text{Ar}$ ) studies also indicate that the northern Gawler Craton was actively exhumed at ca. 1480–1440 Ma, the same time as the deposition of the lower Belt Supergroup strata in Laurentia (Reid and Forster, 2021).



**Fig. 2.9:** Paleo–Mesoproterozoic stratigraphy and U–Pb/Lu–Hf detrital zircon results from the Priest River Region (Doughty and Chamberlain, 2008; Box et al., 2020; Lewis et al., 2020; Brennan et al., 2021) and the range of robust  $\epsilon\text{Hf}_t$  values (samples 67DTB19 and 37DTB19) from the underlying ca. 2670 and 1580 Ma igneous basement. The red box indicates the age range (ca. 1.88–1.83 Ga) of Paleoproterozoic tectonism/magmatism within the Clearwater Block (Vervoort et al., 2016). Inferred crustal evolution line ( $\text{Lu}/\text{Hf} = 0.022$ ) for the Clearwater Block indicated. We note that the main ca. 1860 Ma detrital zircon population within the Gold Cup Quartzite contains  $\epsilon\text{Hf}_t$  values of -3 to -8, which are consistent with our suggested crustal evolution trend of the Clearwater Block. DM–Depleted Mantle, LE–Late Extraction, NC–New Crust, CHUR–Chondritic Uniform Reservoir. See Fig. 7 for explanation of mantle evolution lines.

### 2.5.7. Insights into the evolution and configuration of supercontinent Nuna

As previously discussed, with the exception of eastern Australia, notably the Gawler Craton, NAMG magmatism is relatively limited in the global geologic record (Figs. 2.10, 2.11). Thus, the Australian ca. 1600–1500 Ma magmatism is often considered a significant tie-point for locating eastern Australia–East Antarctica against western Laurentia during the early Mesoproterozoic (e.g. Goodge et al., 2017; Kirscher

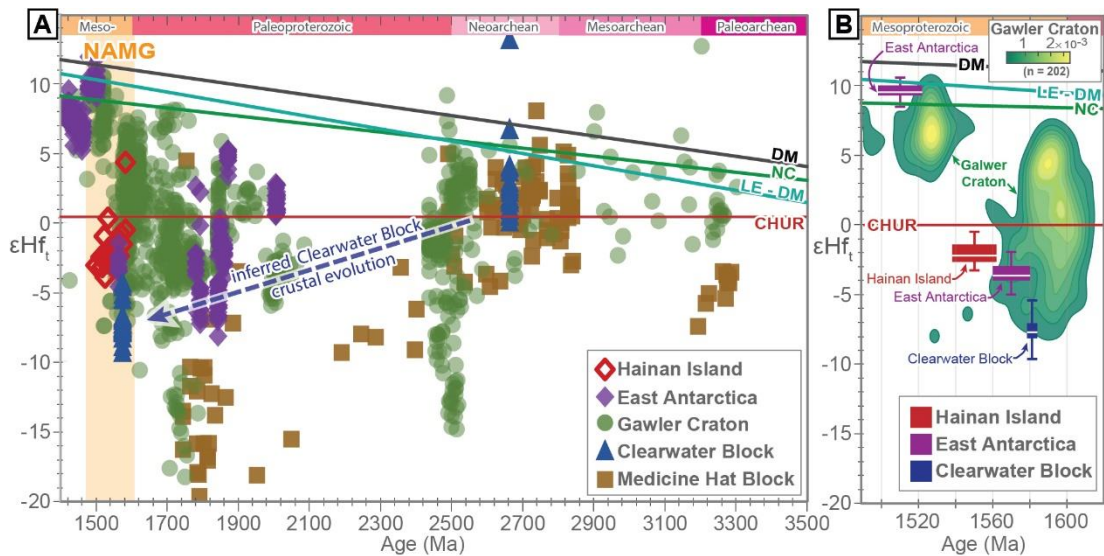
et al., 2019, 2021). In some reconstructions Hainan Island, including all or some of the Cathaysia Block of South China (cf. Yao et al., 2017; Xu et al., 2019; Cawood et al., 2020), is also placed in close proximity to western Laurentia in Nuna. Within southeastern Australia, NAMG-age magmatism is predominately juvenile ( $+\epsilon\text{Hf}_i$ ) and is episodic with two main pulses at ca. 1620–1575 Ma and ca. 1525 Ma, respectively (Fig. 2.10; Reid and Payne, 2017; Chapman et al., 2019; Hartnady et al., 2020). Granitic glacial clasts sampled from East Antarctica also contain a similar, two-pulsed magmatic record of NAMG-age magmatism (Goodge et al., 2017), consistent with East Antarctica being part of the Mawson continent together with the Gawler Craton for most of the Proterozoic (e.g., Payne et al., 2009). However, the ca. 1570 Ma magmatism in East Antarctica is more evolved than the mostly juvenile similar-age magmatism within the Gawler craton (Fig. 2.10). Compared to the East Antarctica constraints, the Laclede Gneiss (sample 37DTB19) is a slightly better age-match for Gawler craton NAMG magmatism but is not an ideal Lu–Hf match due to its more evolved composition.

The ca.  $1551 \pm 12$  Ma felsic magmatism recently recognized within Hainan Island of south China (Xu et al., 2019) is perhaps the worst age-match for magmatism within the Gawler Craton and East Antarctica, as it falls between the two main pulses of magmatism. It is also  $\sim 30$  Myr younger than the Laclede Gneiss but is similar to nearby  $1551 \pm 5$  Ma mafic dykes in Laurentia (Rogers et al., 2018). Admittedly, the ca. 1550 Ma magmatism in Hainan Island does have the least evolved (most juvenile) Lu–Hf values, making it a slightly better Lu–Hf match for the Gawler Craton than other discussed ca. 1580–1550 Ma rocks (the Laclede Gneiss and East Antarctica glacial clasts). Interestingly, northeastern Australia (Mount Isa Inlier; Fig. 2.11B) also contains poorly studied ca. 1550–1500 Ma felsic magmatism (Page and Sun, 1998; Li et al., 2020) that appear to be a better age match with Hainan Island than the Gawler Craton. These ca. 1500 Ma granitoids in northern Australia likely provided the detrital zircon grains found within the ca. 1460–1400 Ma PR1 basin in northwestern Laurentia (Medig et al., 2014).

Genesis models for the prevalent NAMG-age magmatism in southeastern Australia are complex. However, the ca. 1600–1500 Ma magmatism has generally been associated with the final assembly of Australia and East Antarctica with western Laurentia to form Nuna (Betts et al., 2008; Payne et al., 2009; Kirscher et al., 2019). Most models emphasize mantle plume activity that may have modified subduction-

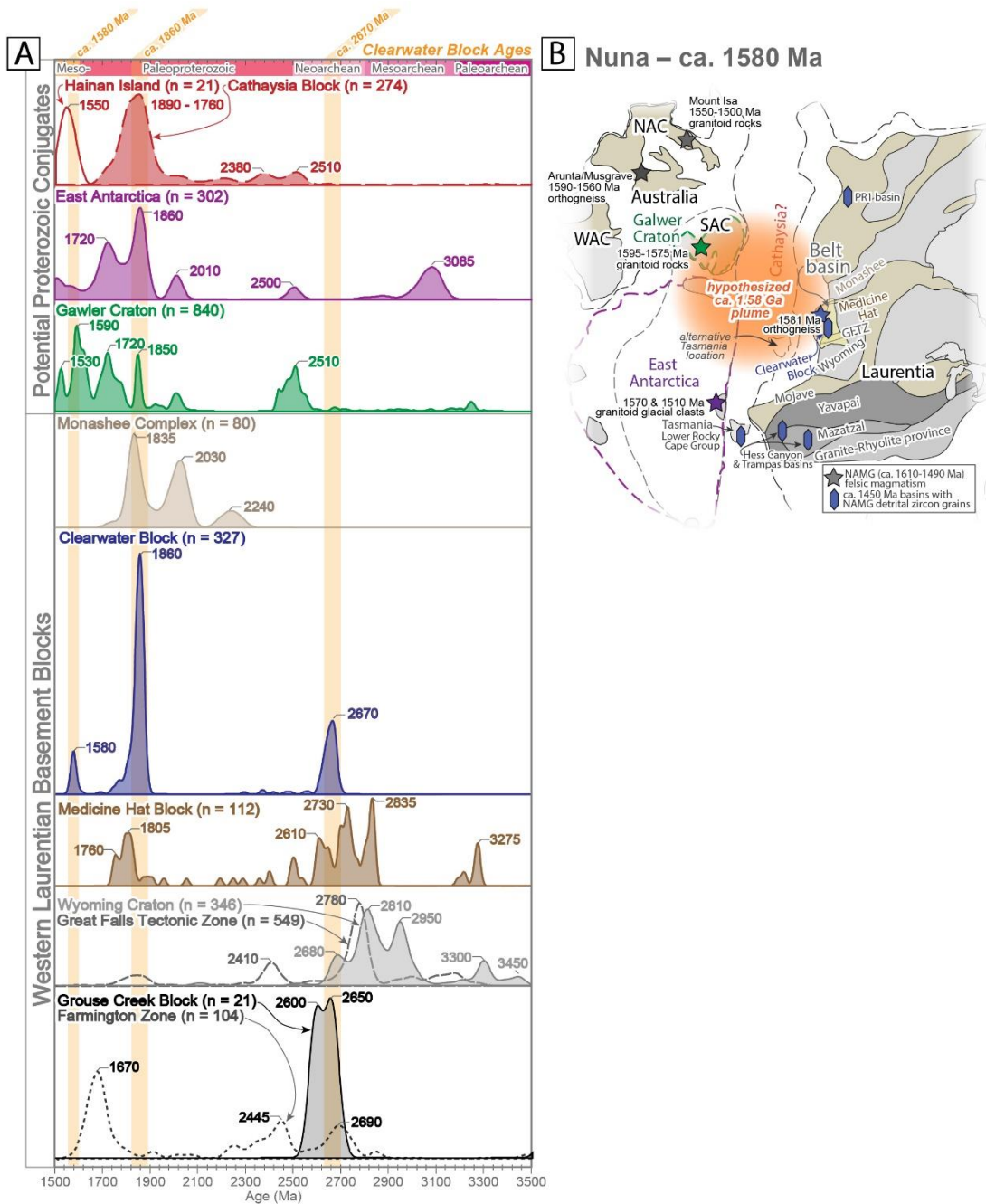


related arc-magmatism, localized back-arc extension and/or continental collision orogenic processes (Betts et al., 2009; Payne et al., 2009; Forbes et al., 2011; Tiddy et al., 2020). Given the absence of a clear subduction-related isotopic signature in the Clearwater Block at this time, we suggest that a mantle plume, perhaps centered beneath southeastern Australia, supplied the heat to melt the (likely) adjacent Archean lower continental crust within the Clearwater Block (Fig. 2.11B), resulting in the generation of the evolved ca. 1580 Ma Laclede Gneiss. This proposed petrogenesis explains the predominantly juvenile NAMG magmatism in southeastern Australia, indicative of a location above the plume head, as well as the plume related geochemistry for the ca. 1590 Ma dykes in western Laurentia (Rogers et al., 2018) and our results for the Laclede Gneiss. This model is also consistent with paleomagnetic reconstructions of Nuna (Fig. 2.11B) which place the Gawler Craton adjacent to the Clearwater Block at ca. 1580 Ma (e.g. Kirscher et al., 2019, 2021; Pisarevsky et al., 2014).



**Fig. 2.10:** A) Individual  $\epsilon\text{Hf}_t$  vs.  $^{207}\text{Pb}/^{206}\text{Pb}$  ages for the Clearwater Block (including the inferred crustal evolution trend), Medicine Hat Block (Gifford et al., 2020) and the suggested Non-Laurentian conjugate blocks of Hainan Island (South China; Xu et al., 2019), East Antarctica (Goodge et al., 2017), and the Gawler Craton of southeastern Australia (Reid and Payne, 2017 and references therein; Hartnady et al., 2020). Inferred crustal evolution line ( $\text{Lu}/\text{Hf} = 0.022$ ) for the Clearwater Block indicated. B)  $\epsilon\text{Hf}_t$  vs.  $^{207}\text{Pb}/^{206}\text{Pb}$  values for ca. 1610–1490 Ma, North American Magmatic Gap magmatism. For comparison, the abundant magmatism of the Gawler Craton is plotted in a two-dimensional kernel density estimate plot with 1  $\epsilon\text{Hf}$  and 5 Ma bandwidths (Spencer et al., 2020) and individual NAMG-age granites sampled in Hainan Island (Xu et al., 2019), East Antarctica (Goodge et al., 2017) and the Clearwater Block (this work) shown in box and whisker plots, where the width of the box corresponds to the  $^{207}\text{Pb}/^{206}\text{Pb}$  age uncertainty. Note that the evolved ca. 1580–1550 Ma magmatism in the Clearwater Block, East Antarctica, and Hainan Island are not a particularly good match for similar-age, juvenile magmatism within the Gawler Craton. DM–Depleted Mantle, LE–Late Extraction, NC–New Crust, CHUR–Chondritic Uniform Reservoir. See Fig. 2.7 for explanation of mantle evolution lines.





**Fig. 2.11:** A) Timing of Paleoproterozoic to early Mesoproterozoic magmatism for the main Archean blocks of western Laurentia and proposed conjugate blocks within Nuna shown in kernel density estimate plots with 25 Ma bandwidths. The main ca. 2670, 1860 and 1580 Ma ages of Clearwater Block magmatism are indicated in orange. Age peak labels are rounded to the nearest 5 Ma. Sources are compiled in the supplementary but include: Grouse Creek (Link et al., 2017); Farmington Zone (Mueller et al., 2011); Great Falls Tectonic Zone (Mueller et al., 2002; Foster et al., 2006; Gifford et al., 2018); Wyoming Craton (Frost and Fanning, 2006; Frost et al., 2006; 2017; Mueller et al., 2010); Medicine Hat Block (Gifford et al., 2020); Clearwater Block (Vervoort et al., 2016; this work); Monashee complex (Crowley et al., 1999); Gawler Craton (Reid and Payne, 2017 and references within; Hartnady et al., 2020); East Antarctica (Goodge and Fanning, 2016; Goodge et al., 2017); the Cathaysia Block of South China (Zhang et al., 2021 and references within); and Hainan Island of South China (Xu et al., 2019). B) Paleogeographic reconstruction of Nuna (after Goodge et al., 2017; Kirscher et al., 2019, 2021; Pisarevsky et al., 2014) with the extent of our proposed ca. 1580 Ma plume, NAMG-age (North American Magmatic Gap) felsic magmatism (indicated by star symbols), and Laurentian basins containing NAMG-age detrital zircon (indicated by tetragonal crystal symbol; for Hess Canyon and Trampas basins, see Doe et al., 2012; 2013; Jones et al., 2015, for PR1 basin see Medig et al., 2014) shown. A more detailed map of western Laurentia's basement framework is shown in Fig. 2.1A.

## 2.6. Conclusions

This work presents zircon U–Pb geochronological and comprehensive isotopic (Lu–Hf,  $\delta^{18}\text{O}$ , and trace element) study of Mesoproterozoic and older crystalline basement rocks within the Clearwater Block of western Laurentia, a key Proterozoic supercontinent tie-point that contains a unique record of NAMG-age magmatism. In addition to providing robust source terrane characterization for utilization by future provenance studies, the main findings from our study of the Clearwater Block crystalline basement are the following:

1) The juvenile Lu–Hf, mantle-like  $\delta^{18}\text{O}$ , and trace element results from the oldest, ca. 2670 Ma, granitic basement rocks indicate that Neoproterozoic crustal growth within the Clearwater Block did not involve significant reworking or assimilation of older crust, nor that it requires modern-style subduction processes. The isotopic record therefore does not necessitate that of the Clearwater Block was built onto the edge of the Mesoproterozoic Medicine Hat Block (or the Wyoming Craton either) in a modern plate-tectonic arc setting. Rather, we suggest that ca. 2670 Ma rocks within the Clearwater Block represent either juvenile crustal growth through direct mantle extraction without subduction-related processes, or remelting of an older Mesoproterozoic mafic lower crust.

2) The Neoproterozoic crust in the Priest River complex (northern Clearwater Block) was metamorphosed and melted (see Vervoort et al., 2016) during a ca. 1860 Ma event. We suggest this event likely records collision related to the assembly of Laurentia. Results from the Priest River complex are consistent with models involving north-dipping subduction within the Vulcan Structure.

3) The new  $1581 \pm 3$  Ma crystallization age for the Laclede Gneiss supports prior assertions of NAMG-age magmatism within the Priest River complex based on discordia regression-age calculations (e.g. Evans and Fischer, 1986; Doughty et al., 1998). The evolved Lu–Hf, mantle-like  $\delta^{18}\text{O}$ , and trace element zircon values from the Laclede Gneiss are difficult to reconcile with a subduction or rift origin. Consequently, we favour a plume genesis, as is suggested for the nearby Laurentian  $1590 \pm 5$  Mammoth dykes (Rogers et al., 2018). Similar-age magmatism in southeastern Australia also has plume-related genesis models (e.g. Reid and Payne 2017; Chapman et al., 2019). The proposed Lu–Hf ‘reworking array’ and O- isotope values from the Clearwater Block indicate that reworking of a nearby Neoproterozoic

lower continental crustal reservoir is the likely origin for the Laclede Gneiss. Consequently, we propose that the Laclede Gneiss is a crustal-derived, distal manifestation of a ca. 1580 Ma plume event beneath southeastern Australia.

4) The new geochronology and isotopic constraints from the Clearwater Block lend further support to interpretations (e.g. Brennan et al., 2021) that the overlying Gold Cup Quartzite and lower Belt Supergroup strata record a provenance shift from proximal Laurentian, to non-Laurentian sources sometime between ca. 1800 and 1480 Ma. This provenance shift likely corresponds to the final assembly of supercontinent Nuna/Columbia during this interval.

5) The Clearwater complex may contain NAMG-mafic magmatism. However, if present, the record of this event was overprinted by ca. 1410–1350 Ma hydrothermal metamorphism, likely associated with magmatism and rifting within the Mesoproterozoic Belt Basin.

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# **Chapter 3: Detrital zircon U–Pb and Hf signatures of Paleo-Mesoproterozoic strata in the Priest River region, northwestern USA: A record of Laurentia assembly and Nuna tenure**

## **Abstract**

Rocks of the Gold Cup Quartzite, Belt Supergroup and Deer Trail Group crop out in the Priest River region of northeastern Washington and northern Idaho (USA). As these sequences represent the westernmost exposures of Paleoproterozoic–Mesoproterozoic strata between New Mexico (USA) and the northern Yukon (Canada), they are key to understanding the assembly of Laurentia and evaluating the duration and configuration of the supercontinent Nuna/Columbia. Here, we present detrital zircon U–Pb and Lu–Hf isotope data from these sequences. The results indicate that the <1.73 Ga Gold Cup Quartzite contains similar detrital zircon components to other widespread late Paleoproterozoic units that record final assembly of the main Archean blocks of western Laurentia; as such, they do not require any non-Laurentian source. Prominent ca. 1.7–1.5 Ga, isotopically juvenile zircon grains form the main component of the overlying ca. 1.47 Ga Lower Belt Supergroup. Some of these detrital zircon components lack a Laurentian source and are instead consistent with derivation from the Gawler Craton in southeastern Australia. The >1.38 Ga upper Belt Supergroup strata are structurally overlain by <1.3 Ga Deer Trail Group rocks that contain detrital zircon components consistent with mixed Laurentian sources. The paucity of any syndepositional volcanic detrital zircon in Deer Trail Group strata along with its fine siliciclastic and carbonate nature suggest deposition within a tectonically-quiet basin coeval with breakup of Nuna.

## **3.1. Introduction**

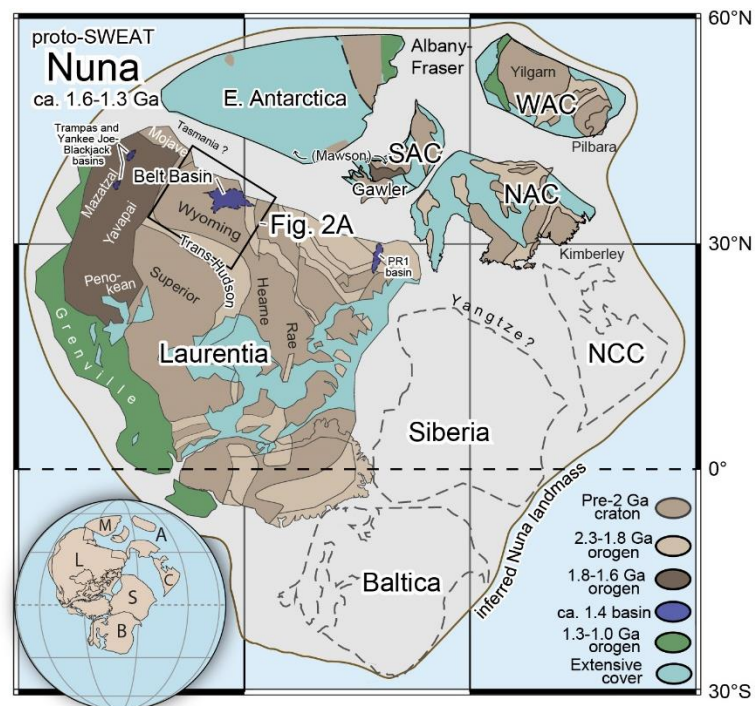
The cyclic amalgamation and dispersal of continents has punctuated Earth's history for at least the past two billion years in a process known as the supercontinent cycle (e.g. Li et al., 2019; Mitchell et al., 2021; Nance et al., 2014). Similarities in stratigraphy, sedimentary provenance and crustal evolution between western Laurentia

(all directions expressed in modern-day coordinates), East Antarctica, and Australia have been used to argue that they were probably connected during most of the Proterozoic, a configuration referred to as the southwest-U.S.–East Antarctic, or SWEAT, connection (Moores, 1991).

Early models suggested the existence of a SWEAT-like connection from as early as 1.9 Ga, and that this configuration may have remained relatively stable until eventual breakup during the late Neoproterozoic (Moores, 1991; Dalziel, 1991; Hoffman, 1991). However, paleomagnetic-based paleogeographic reconstructions indicate that although a SWEAT-like configuration is permissible for most of the early Mesoproterozoic and Neoproterozoic, Australia and Laurentia appear to have been widely separated at ca. 1.25 Ga (Kirscher et al., 2019; Pisarevsky et al., 2014a,b; Kirscher et al., 2020). This suggests that some of the long recognized geological similarities between western Laurentia and Australia–East Antarctica (Moores, 1991; Bell and Jefferson, 1987) may instead provide evidence for the existence of two separate, but perhaps similarly configured, supercontinents—an older, predominately late Paleoproterozoic to Mesoproterozoic supercontinent called Nuna (or Columbia), and a younger, predominately Neoproterozoic, supercontinent called Rodinia (Zhao et al., 2002, 2004; Li et al., 2008a; Meert, 2012; Zhang et al., 2012; Kirscher et al., 2019; Li et al., 2019; Pisarevsky et al., 2014a; Kirscher et al., 2020). This SWEAT-like paleogeography during the late Paleoproterozoic to Mesoproterozoic between Australia, Antarctica and western Laurentia has been called a “proto-SWEAT” configuration (Fig. 3.1; Payne et al., 2009; Kirscher et al., 2019; 2020).

The main assembly of Laurentia occurred during ca. 1.92–1.77 Ga amalgamation of the Archean Rae, Slave, Hearne, Wyoming, Superior, and Medicine Hat cratons along the Trans-Hudson orogen and associated structures including the Great Falls Tectonic zone and Vulcan structure (Fig. 3.2A; Eaton et al., 1999; Mueller et al., 2002; Whitmeyer and Karlstrom, 2007). Similar-age ca. 2.1–1.8 Ga orogens also separate many Archean–Paleoproterozoic blocks in Australia, Siberia, North China, Antarctica, Africa and Baltica, which has led to suggestions that global assembly of Nuna occurred by ca. 1.8 Ga (Rogers and Santosh, 2002; Zhao et al., 2002). However, paleomagnetic (Pisarevsky et al., 2014b; Kirscher et al., 2019; 2020 and references therein) and geological (e.g. Betts et al., 2016; Furlanetto et al., 2013; Li et al., 2020; Nordsvan et al., 2018; Volante et al., 2020a; Volante et al., 2020b) constraints have suggested that an ocean existed between northern Australia and northwestern

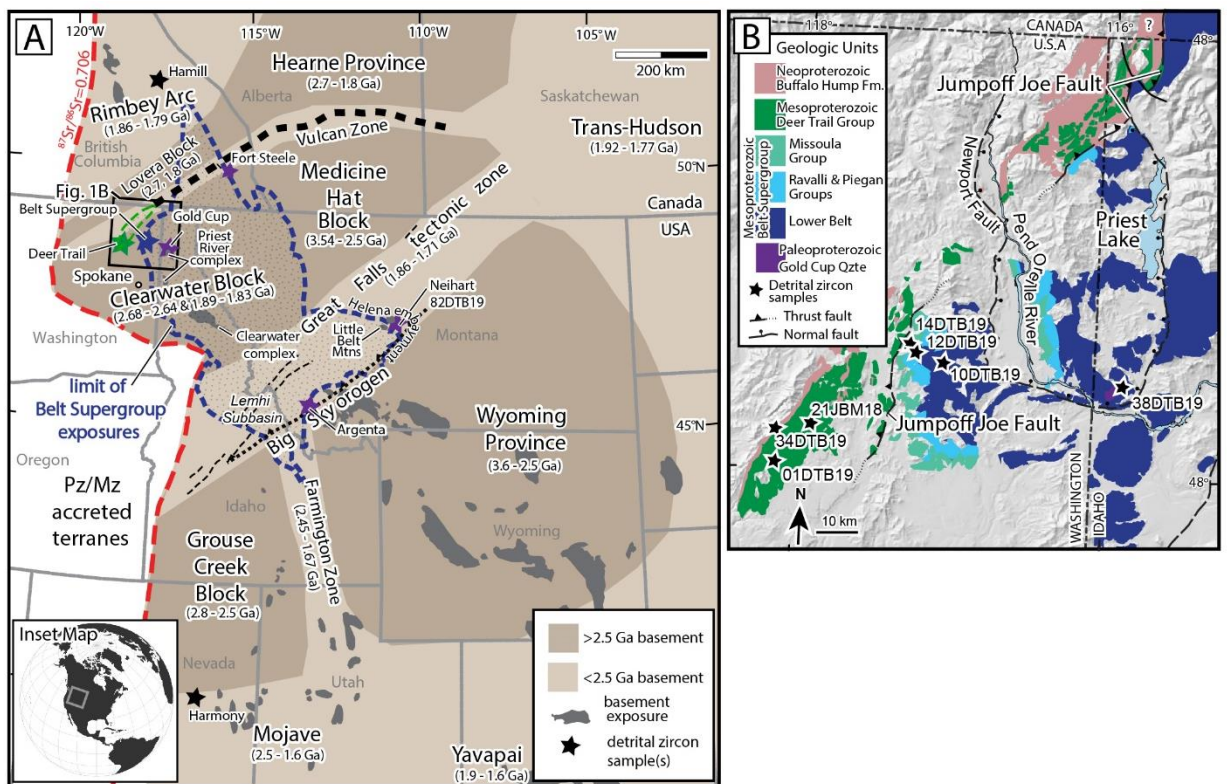
Laurentia from ca. 1.8 to 1.6 Ga. Thus, whether Nuna assembly occurred at ca 1.8 Ga coeval with widespread Trans-Hudson events in Laurentia (e.g. Zhao et al., 2002), or rather involved a more protracted and complex, perhaps two-stage process, lasting from ca. 1.9 to 1.6 Ga (e.g. Kirscher et al., 2019), are debated. In the two-stage model, initial supercontinent amalgamation began with the assembly of a unified Laurentia–Baltica–Siberia landmass by ca. 1.8 Ga, but a final proto-SWEAT configured Nuna was not reached until ca. 1.6 Ga following closure of a small ocean between the Laurentia–Baltica–Siberia and Antarctica–Australia landmasses (Betts et al., 2016; Kirscher et al., 2019; Pourteau et al., 2018; Wang et al., 2020; Kirscher et al., 2020; Nordsvan et al., 2018).



**Fig. 3.1.** Configuration of proto-SWEAT Nuna including its core constituents of Laurentia, Mawson (East Antarctica and Gawler Craton), Australia (NAC–North Australian craton, SAC–South Australia craton, WAC–West Australia craton), North China Craton (NCC), Siberia, and Baltica (adapted from Kirscher et al., 2020). The North China Craton, Siberia, and Baltica are not pertinent to this study and thus only outlines are shown. Paleomagnetic constraints (Kirscher et al., 2020 and references therein) suggest this configuration was stable (albeit underwent counterclockwise rotation) from ca. 1.6–1.3 Ga, and is shown here at ca. 1.3 paleolatitude. Extent of Fig. 3.2A, the Belt Basin, and generally coeval Trampas and Yankee–Joe Blackjack and PR1 basins (Doe et al., 2013; Jones et al., 2015) indicated. Additional abbreviations: L–Laurentia, M–Mawson, A–Australia, C–North China Craton, S–Siberia, B–Baltica.

Regional tectono-stratigraphic evaluation of the proposed two-stage model for Nuna assembly along the Laurentian margin has so far neglected an approximately 3000 km section of the margin (Fig. 3.1) between the PR1 Basin in northwestern Laurentia (Medig et al., 2014; Verbaas et al., 2018), and the Mojave province in southwestern Laurentia (Holland et al., 2015, 2018). The Priest River region (Fig. 3.2) contains one of the most complete records of the Neoproterozoic to Mesoproterozoic tectonic history of the west-central margin of Laurentia and thus offers a unique opportunity to evaluate the timing, duration and credibility of proposed relationships with Australia/Antarctica.

Here, we use existing U–Pb detrital zircon datasets and new U–Pb and Lu–Hf detrital zircon data from three unconformity-bounded sequences in the Priest River region of Laurentia, including the < 1.73 Ga Gold Cup Quartzite, the ca. 1.47–1.38 Belt Supergroup, and the < 1.3 Ga Deer Trail Group. These new data record changes in the provenance of western Laurentia stratigraphy, which can be related to the timing of amalgamation of Laurentia, and the timing of amalgamation then breakup of Nuna.



**Fig. 3.2.** A) General tectonic map of western Laurentia showing Laurentian basement terranes, the limit of Belt Supergroup exposures and geographic areas mentioned in text (after Foster et al., 2006; Whitmeyer and Karlstrom, 2007; Vervoort et al., 2016 and references therein). The extent of regional geologic map in Fig. 3.1B indicated. The  $87\text{Sr}/86\text{Sr} < 0.706$  line approximates the western edge of Laurentia generated during Neoproterozoic rifting (Armstrong et al., 1977; Elison et al., 1990). B) Geologic map of the Priest River region showing the extent of Proterozoic rocks and sample locations (map modified from Miller, 2001; Schuster, 2005 and Lewis et al., 2012).

## **3.2. Geological Setting**

### **3.2.1. The Archean–Paleoproterozoic basement framework of western Laurentia**

In the Priest River and Clearwater complexes (Fig. 3.2A) of northeastern Washington and northern Idaho (USA), exposures of ca. 2.67–2.61 and 1.88–1.84 Ga orthogneisses are considered to belong to the same Archean–Paleoproterozoic Clearwater Block (Vervoort et al., 2016). The Clearwater Block, which represents the westernmost exposures of Laurentian crust beneath the Belt Basin (Foster et al., 2006; Vervoort et al., 2016; Gifford et al., 2020), abuts, or is possibly an extension of, the Medicine Hat Block. The Medicine Hat Block is not exposed, but based on limited borehole and xenolith data is thought to be a mix of Archean (3.1–2.6 Ga) and Paleoproterozoic (ca. 1.81–1.75 Ga) crust (Gifford et al., 2018, 2020; Ross et al., 1991; Villeneuve et al., 1993). Highly evolved ( $\epsilon\text{Hf}t = -8$  to  $-23$ ) ca. 1.8 Ga magmatism within the Medicine Hat Block suggests that the younger Paleoproterozoic period of crustal growth reworked an older Meso- to Neoproterozoic component (Gifford et al., 2020).

The Archean Wyoming Province and Grouse Creek Block are the southern neighbors of Clearwater and Medicine Hat blocks (Fig. 3.2A; Whitmeyer and Karlstrom, 2007). The Wyoming Province consists of temporally and geochemically distinct sub-provinces, including rocks as old as 3.6 Ga and detrital zircon ages as old as 4.0 Ga (Chamberlain et al., 2003; Mueller and Frost, 2006). The Grouse Creek Block in central Idaho contains ca. 2.67–2.50 felsic and ca. 1850 Ma mafic magmatic rocks (Link et al., 2017).

Paleoproterozoic (Trans-Hudson-age; ca. 1.9–1.7 Ga) suture/collisional zones separate the main Archean blocks of western Laurentia. The boundary between the Medicine Hat Block and the Wyoming Province to the south (Mueller et al., 2002) is delineated by the broad Great Falls Tectonic Zone, which records major ca. 1.86–1.73 Ga Paleoproterozoic convergent events including the Great Falls Orogeny and the more temporally and spatially restricted Big Sky Orogeny. Generally, Great Falls Tectonic Zone rocks indicate northwest-dipping subduction, closure of an intervening ocean and eventual amalgamation of the Medicine Hat Block to the Wyoming Province (Mueller et al., 2002; Cheney et al., 2004; Harms et al., 2004; Foster et al., 2006; Alcock et al., 2013; Condit et al., 2015; Gifford et al., 2014; 2018; 2020).

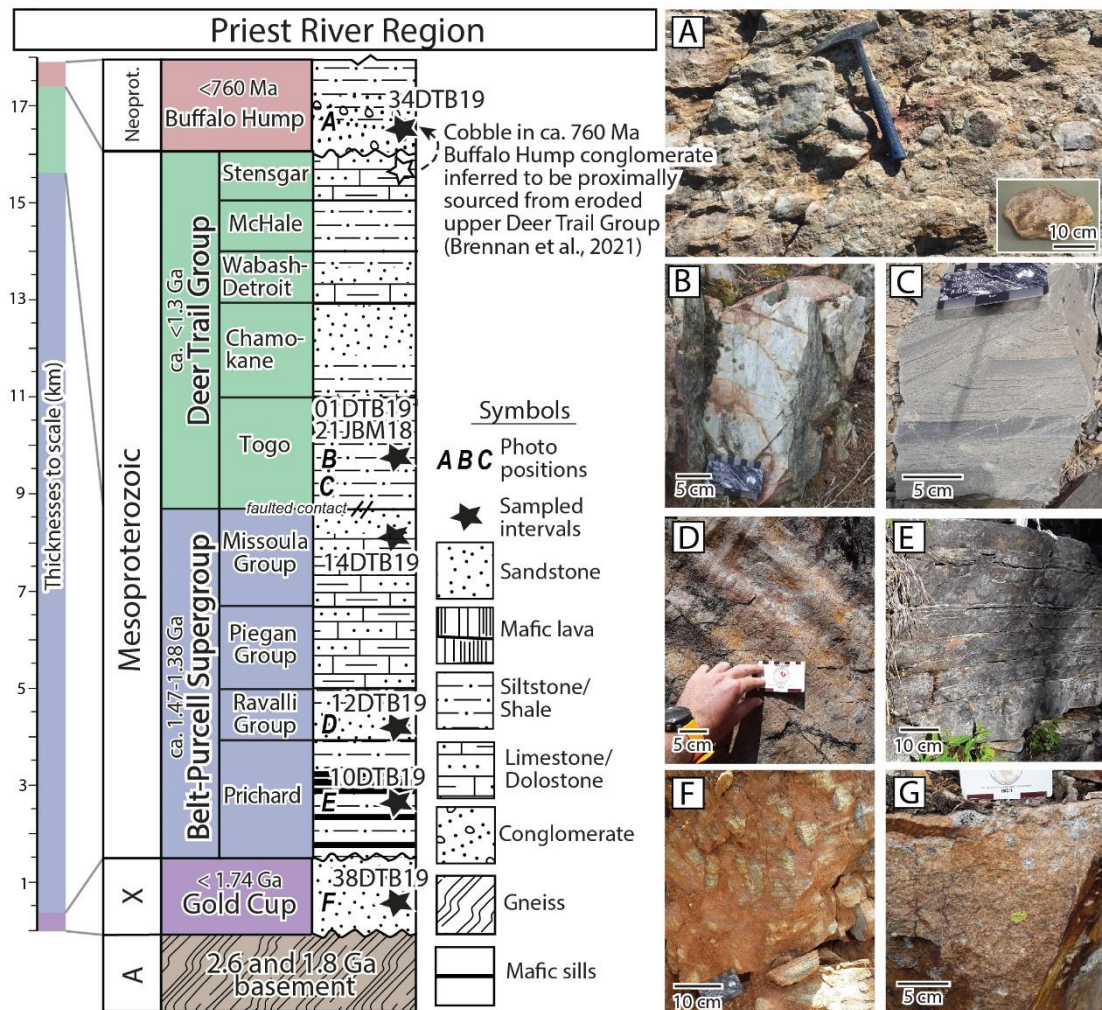
The Vulcan Zone is a Paleoproterozoic collisional belt that records north-dipping subduction between the Archean Medicine Hat–Clearwater block(s) and the southwestern Hearne Province (Eaton et al., 1999; Nieuwenhuis et al., 2014). The Loverna Block, which represents the most southwestern exposures of the Hearne Province, consists of ca. 2.71, 1.82, and 1.78 Ga rocks that are bordered to the north by the ca. 1.85–1.79 Ga Rimbey Arc (Ross et al., 1991).

The Farmington Zone in northeastern Utah contains primarily ca. 2.45 Ga *meta*-igneous rocks and < 2.42 Ga metasedimentary rocks that record upper amphibolite to granulite facies metamorphism at ca. 1.67 Ga (Link et al., 2017; Mueller et al., 2011; Nelson et al., 2011). Although the extent and tectonic significance of the Farmington Zone is uncertain, it likely: (i) separates the Archean Wyoming Province from the Grouse Creek Block to the west, and (ii) records the ca. 1.67 Ga collision of the predominantly late Paleoproterozoic Mojave province with the southern margin of the Grouse Creek–Wyoming provinces (Mueller et al., 2011).

Following assembly of the main Archean blocks of western Laurentia, a long-lived accretionary boundary resulted in the addition of the ca. 1.84–1.70 Ga Mojave province, which comprises a mixture of juvenile and evolved components, and the juvenile ca. 1.80–1.70 and 1.65–1.60 Ga Yavapai and Mazatzal provinces along the southern/ southwestern margin of Laurentia (Bowring and Karlstrom, 1990; Holland et al., 2020). The more evolved isotopic signatures of the Mojave province rocks suggest reworking of a Neoproterozoic crustal component, which may suggest an early association with East Antarctica (Holland et al., 2018; Wooden et al., 2013). Collectively, the Mojave–Yavapai–Mazatzal provinces may record progressive accretion of separate, predominantly juvenile arc systems (Whitmeyer and Karlstrom, 2007), or represent an extensive arc-backarc system that developed along the Australia–Antarctica margin at ca. 1.8 before their accretion to Laurentia after ca. 1.65 Ga (Gibson and Champion, 2019). Detrital zircon (U/Pb and Lu/Hf) data from the ca. 1.79–1.74 Ga Vishnu basin in the Mojave province shows similar detrital zircon ages with coeval Australia–Antarctica strata and igneous sources, supporting an early (ca. 1.8 Ga) Mojave–Australia–Antarctica connection (Holland et al., 2015, 2018). Additionally, recognition of ca. 1.6–1.5 Ga metavolcanic rocks is increasing in the southwestern U.S., including a preliminarily dated  $1505 \pm 6$  Ma ash fall tuff within the Yankee Joe basin (Doe and Daniel, 2019), and the  $1588 \pm 7$  Ma Blue Springs metarhyolite within the nearby Manzano group (Holland et al., 2020).



Several unconformity-bounded Paleo–Mesoproterozoic sequences (Fig. 3.3), including: 1) the Paleoproterozoic Gold Cup Quartzite (Doughty and Chamberlain, 2008); 2) the Mesoproterozoic (ca. 1.47–1.38 Ga) Belt Supergroup and; 3) the ca. <1.37 Ga Deer Trail Group (refined to <1.30 Ga by this work; Ross et al., 1992; Box et al., 2020) overlie basement rocks of the Clearwater Block in the Priest River region. Within the Priest River metamorphic complex, metamorphosed Belt Supergroup strata host a small (<3 km<sup>2</sup>) tectonically interleaved slice of ca. 1.58 Ga orthogneiss (the Laclede Gneiss) whose petrogenesis is poorly understood but represents the only known ca. 1.6–1.5 Ga granitoid in western Laurentia. As such, the Laclede Gneiss is suggested to be a piece of Australia that was stranded during Nuna breakup (Doughty and Chamberlain, 1996, 2008; Lewis et al., 2010; Lewis et al., 2020).



(Previous page) **Fig. 3.3.** Composite stratigraphic diagram of Paleo–Mesoproterozoic rocks from the Priest River region. Approximate stratigraphic thicknesses from Miller et al., (1999) and Lewis et al., (2020), and are only shown to scale on the far left of the diagram. Sampled and photographed intervals indicated. Field photographs include A) A conglomerate bed in the ca. 760 Ma Buffalo Hump Formation, and quartzite cobble (34DTB19) that has a likely eroded Deer Trail Group provenance (Brennan et al., 2021a), B) Fine-grained sandstone bed present in the Togo Formation (basal Deer Trail), C) Laminated mudstone showing soft-sediment deformation that is common in the Togo Formation, D) Ripples in Ravalli Group sandstones (Belt Supergroup), E) Laminated sandstones of the Prichard Formation (Lower Belt Supergroup), and F) Poorly exposed, but coarse-grained sandstone of the Gold Cup Formation. Photograph G is not from the Priest River region but rather shows coarse-grained to pebbly sandstone of the pre-Belt Supergroup Neihart Formation. The Neihart Formation crops out along the eastern edge of the Belt Basin, ~500 km east of the Priest River region (see Fig. 3.2A).

### **3.2.2. Proterozoic sequences in the Priest River region**

#### *3.2.2.1. Paleoproterozoic Gold Cup Quartzite and potential correlatives*

An approximately 100 m-thick coarse pebbly quartzite ('Quartzite of Gold Cup Mountain', hereafter referred to as the Gold Cup Quartzite) unconformably overlies ca. 2.67 Ga Neoproterozoic and ca. 1.87 Ga Paleoproterozoic orthogneisses in the Priest River complex (Fig. 3.2B). Although its upper contact with the overlying lower Belt Supergroup equivalents is mylonitized, it is likely a tectonized unconformity (Fig. 3.3; Doughty et al., 1998; Doughty and Chamberlain, 2008; Lewis et al., 2010; Lewis et al., 2020). Doughty and Chamberlain (2008) analyzed 18 detrital zircon grains from the Gold Cup Quartzite and found ca. 2.65 Ga and ca. 1.8–1.93 Ga age components. Based on these data, Doughty and Chamberlain (2008) favored correlation of the Gold Cup Quartzite with the Paleoproterozoic Neihart Formation that crops out along the eastern margin of the Belt Basin, consistent with the lithostratigraphic correlation proposed by Winston et al. (1989).

The ~ 300 m-thick Neihart Formation (the Neihart Quartzite) crops out in west-central Montana (Fig. 3.2A), and unconformably overlies ca. 1.87 Ga basement of the Great Falls Tectonic Zone (Mueller et al., 2002). The Neihart Formation was part of a widespread fluvial-nearshore sand sheet that was deposited prior to deposition of the Belt Supergroup (Freeman and Winston, 1987; Schieber, 1989). Detrital zircon provenance study indicated that the Neihart Formation contains primarily 1.9–1.7 Ga, and ca. 2.6 Ga detrital zircon components that do not support early correlations with lower Belt Supergroup strata from the main part of the basin further west (Mueller et al., 2016). Rather, these results suggest that deposition of the Neihart Formation occurred sometime after ca. 1.7 Ga, possibly during extensional collapse of the Great Falls Orogen, approximately 200 million years prior to onset of Belt Basin subsidence at ca. 1.47 Ga (Mueller et al., 2016).



Other possibly pre-Belt Supergroup quartzites crop out ~250 km southwest of the Little Belt Mountains, near Argenta Montana, where a ~150 m-thick mature quartzite (the quartzite of Argenta) is overlain by Cambrian strata in a shallow, west-dipping, angular unconformity (Sears et al., 2010). The base of the quartzite of Argenta is not exposed. Approximately 150 km southeast of the Priest River complex, along the western edge of the Clearwater complex, another suggested pre-Beltian quartzite, the Marble Creek Quartzite, unconformably overlies ca. 1.86 Ga orthogneiss and is overlain by Lower Belt Supergroup rocks in a likely faulted relationship (Baldwin et al., 2016).

In southeastern British Columbia, ~500 km northwest of the Little Belt Mountains, quartz arenite of the Fort Steele Formation occurs below Mesoproterozoic Belt Supergroup strata (Hoy, 1992; Ross and Villeneuve, 2003). However, the Fort Steele Formation is much thicker (>2000 m) than the other suggested pre-Belt Supergroup quartzites and shows a more varied composition. Generally, quartzite of the Fort Steele Formation fines upwards into deeper water facies siltstones and mudstones, possibly of the basal Belt Supergroup, and likely records an extensive fluvial system that flowed predominantly to the north-northwest (Hoy, 1992).

#### 3.2.2.2. *Mesoproterozoic Belt–Purcell Supergroup*

The relatively fine-grained siliciclastic and carbonate rocks comprising the thick (up to 18 km) Mesoproterozoic (ca. 1.47–1.38 Ga) Belt Supergroup (known as the Purcell Supergroup in Canada) crop out over a large area (>200,000 km<sup>2</sup>) throughout Montana, Idaho, eastern Washington, and southern British Columbia (Fig. 3.2A; Sears, 2007; Lonn et al., 2020; Mueller et al., 2016). The Belt Basin stratigraphy consists of the Lower Belt, Ravalli, Piegan, and Missoula groups (e.g. Link et al., 1993; Sears, 2007; Winston 2007; Fig. 3.3). In the Priest River region, the Lower Belt Group consists of the Prichard Formation, a >6 km-thick sequence of turbidities intruded by ca. 1469 Ma mafic sills that exhibit textures suggesting they were emplaced into unconsolidated sediments (Anderson and Davis, 1995; Sears et al., 1998; Schandl and Davis, 2000).

The Lower Belt Group grades upwards into the Ravalli Group, which shows different western and eastern facies. In the Priest River region, the western facies of the Ravalli Group consist of the Burke, Revett, and St. Regis formations (Lewis et al., 2020; Miller et al., 1999), which are generally fine-grained quartzites that extend eastward and represent a broad alluvial apron formed by massive sheet floods.

Overlying the Ravalli Group, the Piegan Group includes the Wallace and Helena formations. The Helena Formation consists of carbonates containing stromatolitic horizons. Sand-carbonate cycles that commonly include hummocky cross-stratification characterize the overlying Wallace Formation (Winston, 2007).

The stratigraphically highest, Missoula Group marks a return to dominantly siliciclastic facies with widespread fluvial-alluvial fan deposition (Sears, 2007). A localized  $1401 \pm 6$  Ma tuff (the “Libby tuff”; all uncertainties given at  $2\sigma$ ) occurs  $\sim 1.7$  km from the top of the Missoula Group, and constrains the depositional age of upper Belt Supergroup rocks to ca. 1.4 Ga (Evans et al., 2000). Strata in the southwestern part of the Belt Basin (in the Lemhi Subbasin of east-central Idaho), are termed the Lemhi Group and mostly contain Missoula Group equivalents (Winston et al., 1999; Link et al., 2007). In central Idaho (approximately 80 km west of Argenta, Montana; Fig. 3.2A), Lemhi Group rocks are intruded by ca. 1370 Ma granites and contain detrital zircon grains as young as ca. 1380 Ma (Doughty and Chamberlain 2008; Stewart et al., 2010; Link et al., 2016). Collectively, sedimentary features of Belt Supergroup rocks record their deposition within a vast, intracratonic, epeiric sea or lake. This sea may have opened to the ocean to the northwest, and likely experienced intermittent fluvial, playa, and lacustrine depositional systems (Winston, 2016; Winston and Link, 1993; Pratt, 2001).

Strata of the Lower Belt, Ravalli, and Piegan groups are interpreted to record provenance from both eastern (Laurentian) and western (non-Laurentian) sources (Ross and Villeneuve, 2003; Lewis et al., 2007). Specifically, these rocks contain ca. 1610–1490 Ma “North America Magmatic Gap” (NAMG) detrital zircons, which do not have igneous sources within Laurentia (e.g. Ross and Villeneuve, 2003). Consequently, these NAMG-grains are thought to record provenance from non-Laurentian terranes likely now located in Australia (Ross and Villeneuve, 2003; Fanning and Link, 2003) or possibly Antarctica or South China (e.g. Goodge et al., 2017; Xu et al., 2019; Li et al., 2008b).

Box et al. (2020) contended that NAMG zircon grains within the westernmost Missoula Group suggest a connection to a non-Laurentian source until at least ca. 1390 Ma. In rocks of the Missoula Group to the east of the Priest River Complex, NAMG zircon grains are absent and detrital populations comprise mostly juvenile (positive  $\epsilon_{\text{Hf}}$ ) ca. 1770–1640 Ma zircon grains interpreted to reflect sources from southwestern Laurentia (e.g. Ross and Villeneuve, 2003). Correlative strata in the Lemhi Subbasin

contain similar age-populations, and paleocurrent indicators suggest a source to the south (Stewart et al., 2010; Link et al., 2016). The absence of NAMG grains in upper Belt Supergroup rocks, including the Missoula Group and Lemhi Group strata, is commonly interpreted to suggest a switch to sources in southern Laurentia and reorganization of drainage patterns associated with Nuna rifting (Ross and Villeneuve, 2003) and/or onset of the Picuris orogeny in southern Laurentia (Daniel et al., 2013).

Generally contemporaneous ca. 1.45 Ga Laurentian basin formation is noted ~1,500 km to the north of the Priest River region within the (1460–1420 Ma) PR1 basin (Medig et al., 2014) and ~1,000 km to the south of the Priest River region in the (1475–1450 Ma) Trampas and (1488–1436 Ma) Yankee Joe–Blackjack basins and correlatives (Doe et al., 2013; Daniel et al., 2013; Jones et al., 2015; Fig. 3.1). Collectively, Medig et al. (2014) interpreted the formation of these approximately contemporaneous basins to suggest a ca. 1.47–1.40 Ga period of Nuna extension, likely without full continental separation, along a north–south axis.

#### 3.2.2.3. *Late Mesoproterozoic Deer Trail Group*

An estimated 2.5 km-thick package of argillite, siltite, quartzite and dolomite of the poorly-exposed Deer Trail Group consisting of the Togo, Chamokane, Wabash-Detroit, McHale, and Stensgar formations crops out in northeastern Washington State (Fig. 3.2B; Miller, 2000). The basal Togo Formation consists of >650 m of chiefly grey argillite, with a few coarser quartzose beds within the Togo Formation (Fig. 3.3B). Finer-grained intervals of the Togo Formation contain sub-millimeter- to 10 cm-thick bedding, sometimes graded and/or with erosional bases that range from wavy to parallel except where they are disturbed by soft-sediment deformation (Fig. 3.3C). Miller and Whipple (1989) reported syneresis cracks and fluid escape structures within this unit. The lower contact of the Togo Formation is faulted everywhere (Miller, 2000).

The Chamokane Creek Formation overlies the Togo Formation, and is perhaps the least exposed unit of the Deer Trail Group. The Chamokane Creek Formation is comprised of ~600 m of fine-grained sandstone interbedded with minor pale-grey/green argillite. Overlying the Chamokane Creek Formation is the Wabash-Detroit Formation, which is an approximately 400 m-thick, predominantly dolomitic unit with minor argillite and siltite. Overall, the Wabash-Detroit Formation bears an overall lithological similarity to the Togo Formation, but locally contains stromatolitic beds. The McHale Slate stratigraphically overlies the Wabash-Detroit Formation, and

includes ~370 m of dark-grey, green and lavender-gray argillite and shale/slate. The McHale Slate also includes extensive soft-sediment deformation that may increase in abundance toward the northeast (Miller and Whipple, 1989).

The Stensgar Dolomite is the uppermost unit of the Deer Trail Group and consists of ~250 m of dolomite with sparse algal structures (Fig. 3.3), oolites, and nodular chert. Near the inferred top of the unit, Miller and Whipple (1989) reported probable salt casts. Evans (1987) identified no major unconformities within the Deer Trail Group but documented a significant unconformity beneath the (<760 Ma; Brennan et al., 2021a) overlying Buffalo Hump Formation that locally thins the underlying Stensgar Dolomite.

Deer Trail Group strata crop out only in the hanging-wall of the northwest-dipping Jumpoff Joe Fault (Fig. 3.2B; Miller, 2000). This fault was most recently active as a Mesozoic thrust fault. However, the absence of Deer Trail Group rocks in the footwall (to the east) of this fault, and the presence of reworked cobbles of the Deer Trail Group in the <760 Ma Buffalo Hump Formation strata, suggest that this Mesozoic thrust fault is a reactivated Neoproterozoic normal fault (Miller, 2000; Brennan et al., 2021a).

*3.2.2.3.1. Detrital zircon provenance.* Initially, the Deer Trail Group was correlated with the upper Piegan Group and lower Missoula Group of the Belt Supergroup, that only crop out east (in the footwall) of the Jumpoff Joe Fault based on similarities in sedimentary facies (Fig. 3.2B, 3; Miller and Whipple, 1989). However, subsequent detrital zircon provenance analysis (Box et al., 2020) suggested a  $1364 \pm 35$  Ma maximum depositional age (MDA) for the lower Deer Trail Group (Togo Formation), indicating that the entire Deer Trail Group is likely younger than the upper Belt Supergroup. The remainder of the detrital zircon ages within the Deer Trail Group consist primarily of ca. 1.9 to 1.6 Ga and ca. 1.4 Ga age components (Box et al., 2020).

The ~500 m-thick, interbedded siltstone and coarse (locally conglomeratic) quartzite of the Buffalo Hump Formation unconformably overlies upper Deer Trail Group rocks (Fig. 3.3; Evans, 1987; Brennan et al., 2021a). Sandstones from the Buffalo Hump Formation are dominated by ca. 1.2–1.0 Ga detrital zircon grains (Box et al., 2020), but contain a small population of ca. 760 Ma detrital zircon grains, indicating deposition during rifting related to early Rodinia breakup (Brennan et al., 2021a). Quartzite cobbles from conglomerate channels within the Buffalo Hump Formation have a proximal upper Deer Trail Group source based on their large size

and similarity of detrital zircon ages (Box et al., 2020; Brennan et al., 2021a). This suggests removal of an unknown thickness of Deer Trail Group strata during the late Tonian–Ediacaran breakup of Rodinia.

The late Tonian Buffalo Hump Formation is unconformably overlain by <720 Ma Windermere Supergroup correlative strata (Miller, 2000; Brennan et al., 2021a). Cryogenian and Ediacaran rocks of the Windermere Supergroup are widespread in this region of the Cordilleran margin (e.g. Lund et al., 2010; Yonkee et al., 2014; Brennan et al., 2020) and record “Snowball Earth” glacial intervals and poorly-dated rift-related magmatism (Lund and Cheney, 2016).

### **3.2.3 The East Kootenay event: Termination of Belt Supergroup deposition**

A variety of geo- and thermochronometers applied to Belt Supergroup rocks from along the western extent of the Belt Basin record an enigmatic ca. 1.38–1.3 Ga tectono-thermal event, which is commonly associated with Pb–Zn–Ag and Co–Cu ± Au strata-hosted mineralization (Doughty and Chamberlain, 1996; McFarlane and Pattison, 2000; Zirakparvar et al., 2010; Nesheim et al., 2012; McFarlane and Corfu, 2015; Pattison and Seitz, 2012; Slack et al., 2020; Bookstrom et al., 2016). Early work in southern British Columbia suggested that Belt Supergroup strata must have undergone a period of tilting, gentle folding, erosion, and bimodal intrusion prior to deposition of unconformably overlying late Neoproterozoic and Cambrian strata. Some workers considered these observations to represent an East Kootenay “orogeny”, reflecting a (oblique?) collision between Laurentia and a western terrane that terminated deposition of the Belt Supergroup at ca. 1.38–1.30 Ga (McFarlane and Corfu, 2015; McMechan and Price, 1982). Conversely, others have suggested that the ca. 1.38–1.30 Ga East Kootenay event was rather a period of basin rifting, renewed subsidence, and bimodal magmatism that shortly preceded the end of deposition within the Belt Basin (Doughty and Chamberlain, 1996; Pattison and Seitz, 2012).

### **3.3. Field relations and sample descriptions**

To evaluate correlations and provenance interpretations of Paleo- to Mesoproterozoic strata in the Priest River region, existing detrital zircon U–Pb datasets were compiled and additional samples were collected from target stratigraphic intervals for further detrital zircon analysis (U–Pb and Lu–Hf isotopes; Table 1). The stratigraphically-lowest sampled interval was from the core of the Priest River

Complex (Lewis et al., 2020). This sample (38DTB19; Fig. 3.3F) was collected from poorly-exposed, coarse-grained, locally pebbly, Gold Cup Quartzite on the east side of Gold Cup Mountain. Here, the Gold Cup Quartzite overlies Archean igneous basement rocks and is approximately 120 m thick. In turn, a probable unconformity separates the Gold Cup Quartzite from overlying metamorphosed Lower Belt Supergroup strata (Miller et al., 1999; Doughty et al., 1999; Doughty and Chamberlain, 2008; Lewis et al., 2020). To evaluate the regional extent of potential pre-Belt Supergroup units, approximately 500 km east of the Priest River region within the Little Belt Mountains, a sample (82DTB19) of the Neihart Quartzite, a potentially Gold Cup Quartzite correlative (e.g. Doughty and Chamberlain, 2008; Lonn et al., 2020; Mueller et al., 2016), was collected from shallowly-dipping, coarse-grained to pebbly, thickly-bedded, cross-stratified quartzite (quartz arenite) (Fig. 3.3G).

West of the Newport Fault, a relatively intact stratigraphic section of Belt Supergroup rocks dips to the west and is offset by small-scale faults (Lewis et al., 2020). These rocks progress up section to the west (Miller, 2000), where upper Belt Supergroup rocks (or perhaps lowermost Deer Trail Group equivalent rocks; Box et al., 2020) are unconformably overlain by lower Paleozoic strata. In stratigraphic order, a fine-grained sandstone from the Lower Belt Supergroup (Prichard Formation; Fig. 3.3E; 10DTB19) was sampled. Approximately 8 km west of the Prichard Formation sample, medium-grained quartzite of the Revett Formation of the Ravalli Group, was sampled (Fig. 3.3D; 12DTB19). Approximately 3 km southwest of the Revett Formation sample, and ~800 m below the overlying sub-Cambrian unconformity that likely removed most Deer Trail Group strata here (Box et al., 2020; Brennan et al., 2021a), a fine-grained cross-laminated sandstone bed (sample 14DTB19) was sampled from within the predominately argillite sequence of the Snowslip Formation of the Missoula Group.

Deer Trail Group rocks are only mapped west (in the hanging wall) of the Jumpoff Joe Fault (Miller, 2000; Fig. 3.2B). Approximately 20 km southwest of the sampled Belt Supergroup section, an extensively faulted but relatively complete stratigraphic section of the Deer Trail Group is present. In this region, two samples (21JBM18 and 01DTB19; Fig. 3.3B) were collected (~15 km apart but broadly along strike) from the coarsest, medium-grained quartzite interval of the basal Deer Trail Group (Togo Formation) based off the mapping of Evans (1987) and Campbell and Raup (1964).

We also report new Lu–Hf detrital zircon data from a quartzite cobble within the ca. 760 Ma Buffalo Hump Formation that unconformably overlies upper Deer Trail Group. Brennan et al., (2021a) reported the U–Pb ages of the same detrital zircon grains from these cobbles (34DTB19) that are very similar in age to those in upper Deer Trail Group rocks (Box et al., 2020). This suggests that the cobbles were likely reworked from now eroded upper Deer Trail Group, probably during Neoproterozoic rifting (Brennan et al., 2021a).

**Table 3.1**

Paleo-Mesoproterozoic compiled and original detrital zircon samples from the Priest River region & potential Laurentian correlatives.

Geologic Unit	Sample number	Location		Analysis method	Isotopic data	Reference	Figure shown
		Latitude (°)	Longitude (°)				
<u>Now eroded (?) Mesoproterozoic Deer Trail Group</u>							
Buffalo Hump conglomerate	34DTB19	48.1312	–117.9977	LA-SS-ICPMS	U-Pb, Lu-Hf	Brennan et al. (2021a) ; this work	Fig. 7; Fig. 8A*; Fig. 9A
<u>Mesoproterozoic Deer Trail Group</u>							
Wabash-Detroit	07SB303A	48.1654	–117.9185	LA-ICPMS	U-Pb	Box et al. (2020)	Fig. 8A*
Wabash-Detroit	07SB302A	48.1596	–117.9264	LA-ICPMS	U-Pb	Box et al. (2020)	Fig. 8A*
Togo Formation	Togo3	48.1526	–117.9087	LA-ICPMS	U-Pb	Box et al. (2020)	Fig. 8A*
Togo Formation	01DTB19	48.0488	–118.0775	LA-SS-ICPMS	U-Pb, Lu-Hf	this work	Fig. 6, 8A*, Fig. 9A
Togo Formation	21JBM18	48.1619	–117.9272	LA-ICPMS	U-Pb	this work	Fig. 6, 8A*
<u>Mesoproterozoic Belt Supergroup</u>							
Snowslip	14DTB19	48.2713	–117.6567	LA-SS-ICPMS	U-Pb, Lu-Hf	this work	discussed in text
Snowslip	08SB203A	48.2740	–117.6573	LA-ICPMS	U-Pb	Box et al. (2020)	Fig. 8A*
Revelt	12DTB19	48.2909	–117.6284	LA-SS-ICPMS	U-Pb, Lu-Hf	this work	Fig. 5, 8A*, Fig. 9B, 13*
Revelt	07SB306A	48.1247	–117.5834	LA-ICPMS	U-Pb	Box et al. (2020)	Fig. 8A*
Prichard	10DTB19	48.2958	–117.5196	LA-SS-ICPMS	U-Pb, Lu-Hf	this work	Fig. 5, 8A*, Fig. 9B, 13*
Prichard	04RL213	47.4796	–116.2178	LA-ICPMS	U-Pb	Lewis et al., 2010	Fig. 8A*
Prichard	04RL175	48.3747	–116.1934	LA-ICPMS	U-Pb	Lewis et al., 2010	Fig. 8A*
<u>Paleoproterozoic</u>							
<u>Quartzites</u>							
Gold Cup Quartzite	GC-01-03	48.1772	–116.8646	SHRIMP	U-Pb	Doughty and Chamberlain (2008)	Fig. 8A*; Fig. 10*
Gold Cup Quartzite	38DTB19	48.1908	–116.7945	LA-SS-ICPMS	U-Pb, Lu-Hf	this work	Fig. 4; Fig. 8A*; Fig. 9; Fig. 10*, 11*
Neihart Formation	84DTB19	46.9077	–110.6898	LA-SS-ICPMS	U-Pb, Lu-Hf	this work	Fig. 4; Fig. 8A*; Fig. 9; Fig. 10*, 11*
Neihart Formation	VN-98-01	Stop 1-2 of Winston et al., 1989		SHRIMP	U-Pb	Ross and Villeneuve (2003)	Fig. 10*
Neihart Formation	NQ98-2	46.9130	–110.6983	SHRIMP	U-Pb	Mueller et al. (2016)	Fig. 10*
Quartzite of Argenta	07PL7	45.2977	–112.9126	LA-ICPMS	U-Pb	(Sears et al. (2010)	Fig. 10
Fort Steele Formation	RAR-98-FS	49.74088	–115.74255	SHRIMP	U-Pb	Ross and Villeneuve (2003)	Fig. 10

Note: All Locations converted to datum WGS84.

\*data shown compiled with other correlative units

Abbreviations: LA-Laser ablation, SS-Split stream, ICPMS-Inductively Coupled Plasma Mass spectrometry, SHRIMP-Sensitive High Resolution Ion Micro Probe

## 3.4. Methods

### 3.4.1 Detrital zircon analysis

Samples for detrital zircon analysis were collected and separated using standard techniques (chipmunk, disc mill or SEL-Frag, Frantz magnetic and heavy liquid separation) at Curtin University, Australia or the University of Wisconsin-Eau Claire, USA. Zircon grains were mounted in epoxy resin, and hand polished to expose grain centers. Cathodoluminescence (CL) or backscatter electron (BSE) images were taken of all zircon grain mounts. All “DTB” samples were imaged using a MIRA3 variable-pressure field-emission scanning electron microscope (SEM) coupled to a CL

detector and analyzed for U–Pb, and Lu–Hf isotopes using split-stream laser ablation inductively coupled mass spectroscopy (ICPMS) at the GEOHistory Facility in the John de Laeter Centre at Curtin University, Perth Australia. U–Pb isotopes were measured on an Agilent 7700 s quadrupole ICP-MS for age determination, whereas Lu–Hf isotopes were measured on a Nu Instruments Plasma II MC-ICP-MS to further constrain the provenance and character of the source terranes. Following analysis, data reduction was performed using Iolite v.3.6 software (Paton et al., 2011).

Backscattered electron (BSE) images of sample 21JBM18 were collected at University of Wisconsin – Eau Claire using a Hitachi S-3400 N Variable Pressure SEM and then analyzed for U–Pb isotopes at the University of Arizona Laserchron center with a Photon Machine Analyte G2 Excimer laser paired with a Thermo Element2 single-collector ICPMS. Following analysis, data reduction for 21JBM18 was performed with an Arizona Laserchron python decoding routine and an Excel spreadsheet (E2agecalc).

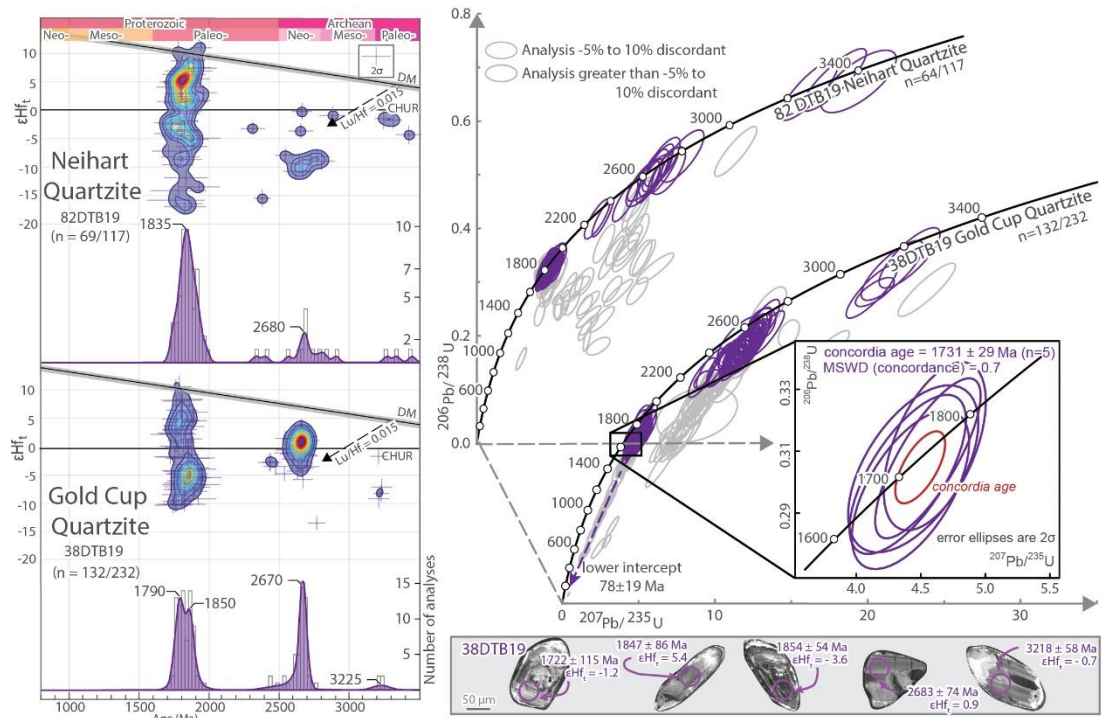
Detrital zircon age spectra are shown in kernel density estimate (KDE) plots generated with DensityPotter 8.5 (Vermeesch, 2012) with 25 Myr band and bin widths. Peak age labels were rounded to the nearest 5 Ma. For zircon grains with ages younger than 1500 Ma, the  $^{206}\text{Pb}/^{238}\text{U}$  age was used, whereas for zircons older than 1500 Ma the  $^{207}\text{Pb}/^{206}\text{Pb}$  age was used (Spencer et al., 2016). For samples, 10DTB19 and 12DTB19 which have depositional ages of ca. 1470 Ma, the  $^{207}\text{Pb}/^{206}\text{Pb}$  ages are plotted. Only analyses with discordance (calculated as  $(^{207}\text{Pb}/^{206}\text{Pb}) - (^{238}\text{U}/^{206}\text{Pb}) / (^{207}\text{Pb}/^{206}\text{Pb}) < 10\%$  to  $-5\%$ ) are interpreted as robust and are included in the uni and bivariate KDE plots. Maximum depositional age constraints only use grains that met these criteria and also showed  $^{206}\text{Pb}/^{207}\text{Pb}$  and  $^{206}\text{Pb}/^{238}\text{U}$  ages that overlapped the concordia within  $2\sigma$  uncertainties (e.g. Spencer et al., 2016). Maximum depositional ages are presented for samples 38DTB19 (Gold Cup Quartzite) and 10DTB19 (Prichard Formation) based on the youngest population with a mean square weighted deviation (MSWD) of  $\sim 1$  (Cou tts et al., 2019; Herriott et al., 2019). Concordia plots and ages were calculated with Isoplot 3.70 (Ludwig, 2008) and IsoplotR (Vermeesch, 2018). The  $\epsilon_{\text{Hf}(t)}$  values calculated for all data used the  $^{176}\text{Lu}$  decay constant =  $1.865 \times 10^{-11} \text{ yr}^{-1}$  proposed by Scherer et al. (2001). Chondritic values are after Bouvier et al. (2008):  $^{176}\text{Hf}/^{177}\text{Hf}$  CHUR = 0.282785 and  $^{176}\text{Lu}/^{177}\text{Hf}$  CHUR = 0.0336, where CHUR is the chondritic uniform reservoir. Depleted mantle values are after Griffin et al. (2002):  $^{176}\text{Hf}/^{177}\text{Hf}$  = 0.28325 and  $^{176}\text{Lu}/^{177}\text{Hf}$  = 0.0384.



Where possible, all published (U–Pb and Lu–Hf) data included were recalculated/reselected using the same criteria.

Bivariate kernel density estimates (2DKDEs) were employed to visualize U–Pb/Lu–Hf data distribution and density with  $1 \text{ } \epsilon\text{Hf}_t$  and 45 Ma bandwidths using the methods and MATLAB code of Spencer et al., (2019). To further assess tectonostratigraphic correlations, and objectively evaluate the relative similarity between datasets, two-dimensional (U–Pb/Lu–Hf) statistical comparison of zircon density distributions utilizing the MATLAB program of Sundell and Saylor (2021) was also employed also using  $1 \text{ } \epsilon\text{Hf}_t$  and 45 Ma bandwidths. See data repository for additional details.

### 3.5. Detrital Zircon U–Pb and Lu–Hf results



**Fig. 3.4.** Detrital zircon U–Pb/Lu–Hf for the Neihart Quartzite (82DTB19) and Gold Cup Quartzite (38DTB19) shown in KDE plots, and Wetherill concordia diagrams. In the Wetherill concordia diagram the purple analyses fit within our concordance limits and are reported in the KDE plots. The grey analyses are outside our concordance limits and are omitted from the KDE plots. The concordia inset shows the five youngest concordant analyses (purple) and their  $1739 \pm 30 \text{ Ma}$  Concordia age (red) from sample 38DTB19. Representative zircon CL-images and individual analysis results are shown on the bottom right for sample 38DTB19. Approximate average crustal evolution ( $\text{Lu/Hf} = 0.015$ ) indicated. Abbreviations: DM–depleted mantle, CHUR–chondritic uniform reservoir.

Detrital zircon grains (Fig. 3.4) from the Gold Cup Quartzite (38DTB19) and Neihart Quartzite (82DTB19) are similar in size, shape and appearance. Zircon grains from both samples are generally 80 to 250  $\mu\text{m}$  long, with median aspect ratios (long

axis/short axis) around ~1.85 but sometimes exceeding 3. In CL imaging grains from both samples show mostly concentric zoning, with some thin (<10  $\mu\text{m}$ ) bright metamorphic rims. U/Pb results show similar age-components with ca. 2.7–2.6 Ga and ca. 1.95–1.7 Ga grains common. Both samples lack zircon grains younger than ca. 1.7 Ga.

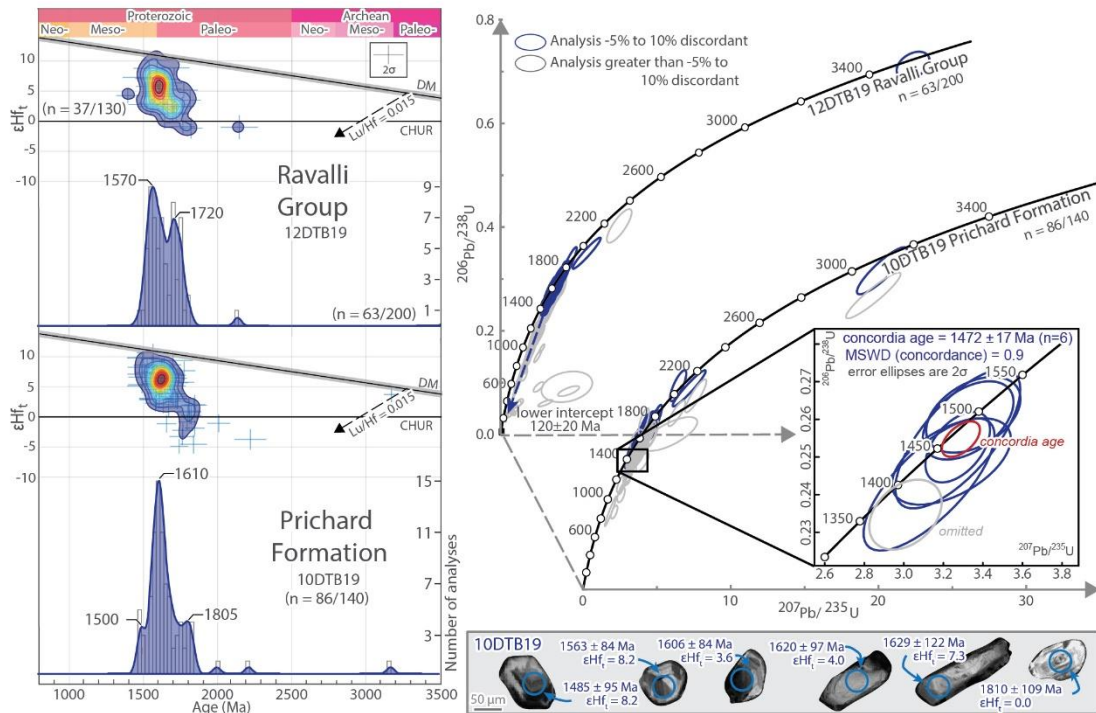
In the Gold Cup Quartzite, a ca. 2.67 Ga age-component yields a tight cluster of  $\epsilon\text{Hf}_t$  values from  $-2$  to  $+3$ , indicating a CHUR-like Neoproterozoic source. The younger detrital component shows main age-peaks at 1850 and 1790 Ma and  $\epsilon\text{Hf}_t$  values that mostly range from  $-10$  to  $+10$ , suggesting similar-aged but isotopically discrete sources. Deposition of the Gold Cup Quartzite occurred after  $1731 \pm 29$  Ma based on a Concordia age constrained by the five youngest concordant analyses with similar  $\epsilon\text{Hf}_t$  values that range from  $-0.9$  to  $3.9$  (Fig. 3.4).

In the Neihart Quartzite, the ca. 2.68 Ga age-component exhibits  $\epsilon\text{Hf}_t$  values ranging from  $-12$  to  $0$ . This is a greater variation in  $\epsilon\text{Hf}_t$  than the similar-aged component in the Gold Cup Quartzite, which contain a tight cluster of  $\epsilon\text{Hf}_t$  values ranging from  $-2$  to  $+3$ . The ca. 1.95–1.7 Ga component in the Neihart Quartzite shows a main age-peak at 1835 Ma and contains a similar range of  $\epsilon\text{Hf}_t$  values (approximately  $-17$  to  $+9$ ) to the temporally equivalent component in the Gold Cup Quartzite.

Detrital zircon grains from the Lower Belt Supergroup (Prichard Formation; 10DTB19), and Ravalli Group (Revett Formation; 12DTB19) are also similar in size, shape and appearance. Zircon grains from both samples are generally 40 to 90  $\mu\text{m}$  long, with median aspect ratios (long axis/short axis) around ~1.9 but sometimes as high as 3, indicating preservation of elongate grains. In CL imaging grains from both samples show mostly faint concentric zoning. U–Pb zircon results show dominant ~1.8–1.5 Ga age-components that contain a significant number of grains with ages within the ca. 1.61–1.49 Ga North American Magmatic Gap (“NAMG”; Ross and Villeneuve, 2003). In the Prichard Formation (Fig. 3.5), the main 1610 Ma age-peak shows a tightly clustered range of  $\epsilon\text{Hf}_t$  values between  $+4$  and  $+9$ . It also contains two shoulder age-peaks, at ca. 1805 Ma and ca. 1500 Ma. The older ca. 1805 age-component shows more evolved  $\epsilon\text{Hf}_t$  values concentrated between  $-6$  to  $+3$ , whereas the younger ca. 1500 Ma age-peak shows juvenile  $\epsilon\text{Hf}_t$  values ( $+6$  to  $+10$ ). A Concordia age of the six youngest (<ca. 1500 Ma) concordant analyses yields an age of  $1472 \pm 17$  Ma (MSWD = 0.9), which overlaps within  $2\sigma$  uncertainty with the  $1468 \pm 2$  and  $1469 \pm 3$  Ma age of mafic sills within the Prichard Formation that likely

intruded unconsolidated sediment and approximate the depositional age (Anderson and Davis, 1995; Sears et al., 1998). These six youngest concordant analyses have similar  $\epsilon\text{Hf}_t$  values from +5.3 to +9.8.

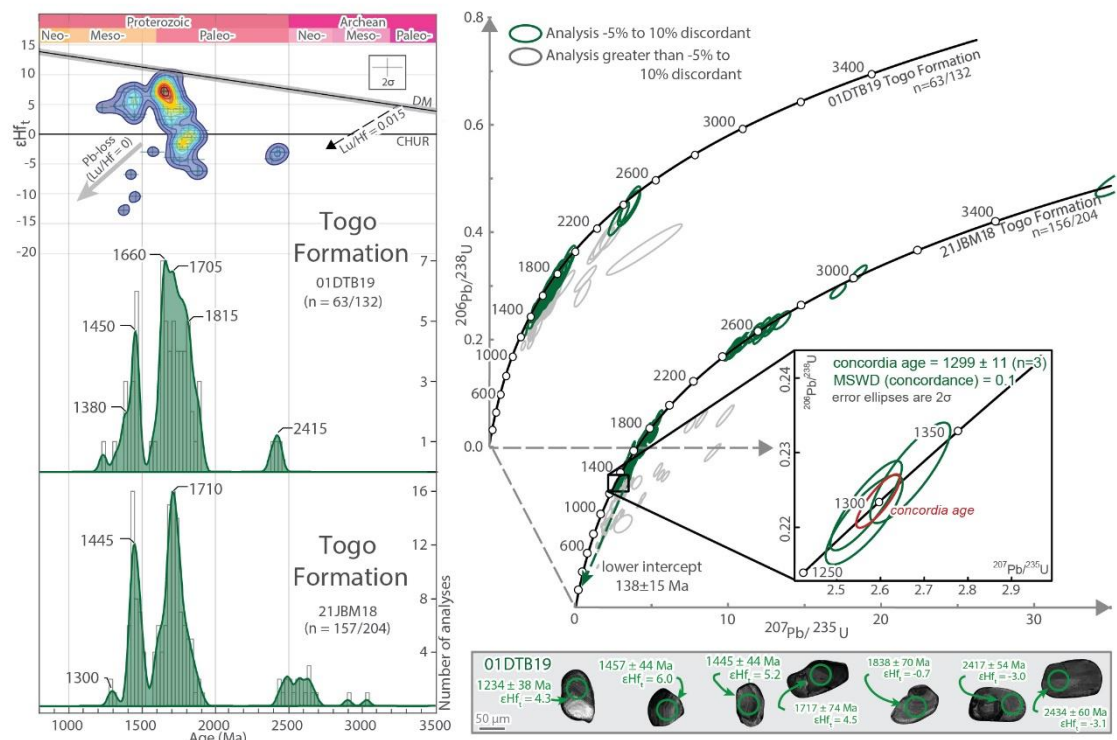
In the Ravalli Group sample (Fig. 3.5), the main age-components are at 1720 Ma and 1570 Ma. Both these age components contain mostly positive  $\epsilon\text{Hf}_t$  values. Detrital zircon grains from the Missoula Group (Snowslip Formation; 14DTB19) experienced significant Cretaceous lead loss and are mostly discordant precluding further interpretation.



**Fig. 3.5.** Detrital zircon U–Pb/Lu–Hf for Belt Supergroup samples from the Prichard Formation (10DTB19) and Ravalli Group (12DTB19) shown in KDE plots, and Wetherill concordia diagrams. The concordia inset shows the five youngest concordant analyses (blue) and their  $1486 \pm 17$  Ma Concordia age (red) from sample 10DTB19. In the Wetherill concordia diagram the blue analyses fit within our concordance limits and are reported in the KDE plots. The grey analyses are outside our concordance limits and are omitted from the KDE plots. Representative zircon CL-images and individual analysis results are shown on the bottom right for sample 10DTB19. Approximate average crustal evolution ( $\text{Lu}/\text{Hf} = 0.015$ ) indicated. Abbreviations: DM–depleted mantle, CHUR–chondritic uniform reservoir.

Detrital zircon grains from the lower Deer Trail Group (Togo Formation; samples 01DTB19 and 21JBM18) are again similar in size, shape and appearance. Zircon grains from both samples are generally 30 to 80  $\mu\text{m}$  long, with median aspect ratios (long axis/short axis) around 1.4, which indicates overall more equant zircon grains than observed within the Belt Supergroup and Gold Cup Quartzite samples. In CL imaging grains from sample 01DTB19 show mostly faint, darker, concentric

zoning. Geochronology results from the Togo Formation indicate an absence of NAMG-age grains, and the presence of age-components younger than ca. 1380 Ma (Fig. 3.6). The Togo Formation samples both show broad-age components from ca. 1.9–1.6 Ga. Sample 21JBM18 shows a single 1710 Ma age-peak, which probably is the amalgamation of several age-populations. In sample 01DTB19, this broad ca. 1.9–1.6 Ga population suggests several discrete subpopulations (Fig. 3.6) with an older age-peak at ca. 1815 Ma characterized by more evolved  $\epsilon_{\text{Hf}t}$  values ( $-7$  to  $+1$ ), and two younger populations at ca. 1705 and 1660 Ma. These younger populations contain more CHUR-like/juvenile  $\epsilon_{\text{Hf}t}$  values around  $+1$  to  $+10$ . The next youngest main age-component at  $\sim 1450$  Ma, is present in both samples. In sample 01DTB19, these grains show mostly juvenile  $\epsilon_{\text{Hf}t}$  values ( $+2$  to  $+8$ ). These main age-components are similar to Missoula and Lemhi Group samples from further east/southeast (e.g. Stewart et al., 2010; Link et al., 2016).



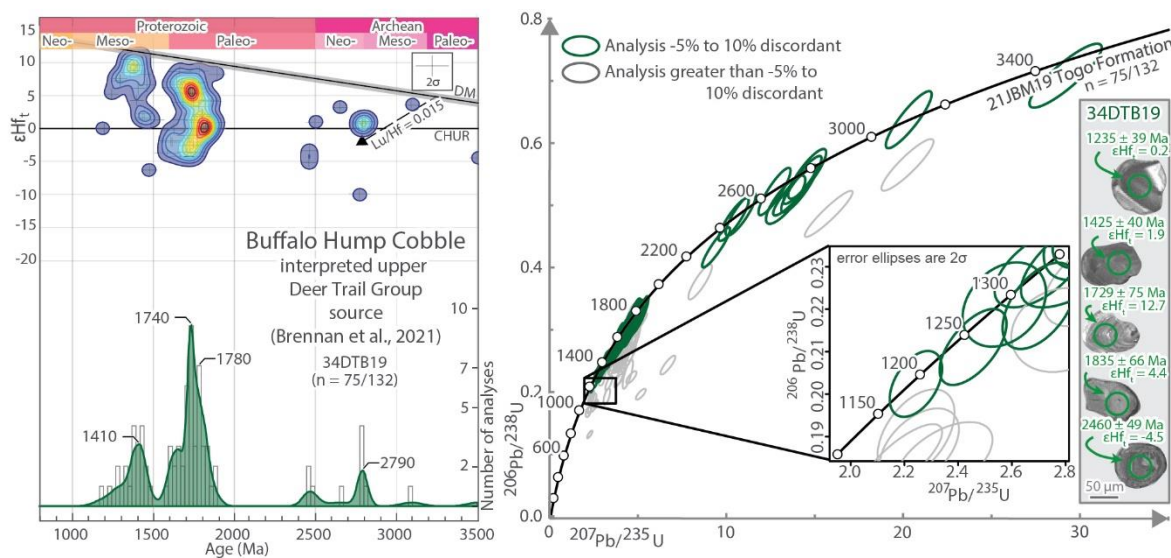
**Fig. 3.6.** Detrital zircon U–Pb/Lu–Hf results for Deer Trail Group samples from the Togo Formation (01DTB19 and 21JBM18) shown in KDE plots, and Wetherill concordia diagrams. In the Wetherill concordia diagram the green analyses fit within our concordance limits and are reported in the KDE plots. The grey analyses are outside our concordance limits and are omitted from the KDE plots. The concordia inset shows the three youngest concordant analyses (green) and their  $1299 \pm 12$  Ma Concordia age (red) from sample 21JBM18, interpreted as the maximum depositional age of the Deer Trail Group. Representative zircon CL-images and individual analysis results are shown on the bottom right for sample 01DTB19. Approximate average crustal evolution ( $\text{Lu}/\text{Hf} = 0.015$ ) indicated. Abbreviations: DM–depleted mantle, CHUR–chondritic uniform reservoir.

We favor a maximum depositional age of  $1299 \pm 12$  Ma for the Deer Trail Group based on a Concordia age of the youngest three overlapping grains in sample 21JBM18 (Fig. 3.6), which was collected from near the base of the group. The other lower Deer Trail Group sample presented here (01DTB19) also contains grains that contain similar  $^{206}\text{Pb}/^{238}\text{U}$  dates of  $1324 \pm 41$  and a youngest single grain date of  $1234 \pm 38$  Ma. A similar sample from the lower Deer Trail Group reported by Box et al., (2020) also contains a single concordant youngest grain at  $1264 \pm 29$  Ma. Therefore, we interpret the Deer Trail Group to be younger than ca. 1.30 Ga.

The U–Pb results from the Buffalo Hump conglomerate cobble (34DTB19) that likely represents eroded upper Deer Trail Group also contain concordant grains with ( $^{206}\text{Pb}/^{238}\text{U}$ ) crystallization dates of  $1188 \pm 34$ , and  $1235 \pm 39$  Ma (Brennan et al., 2021a). These zircon grains are generally larger (lengths of 80 to 200  $\mu\text{m}$ ) than the grains found within the stratigraphically lower Deer Trail Group samples but also contain mostly lower aspect ratios ( $\sim 1.4$ ) and concentric zoning in CL-imaging. Here we present new Lu–Hf data from this cobble (Fig. 3.7) that complement the U–Pb data which indicates the presence of ca. 1.9–1.6 Ga and ca. 1.4 Ga age-components first reported by Brennan et al. (2021a). The ca. 1.9–1.6 Ga component shows a range of  $\epsilon\text{Hf}_t$  values (from  $-5$  to  $+10$ ) suggesting at least two discrete similar-age but different  $\epsilon\text{Hf}_t$  populations. The youngest single grain analysis date of  $1188 \pm 34$  Ma from the Buffalo Hump Cobble overlaps with the youngest concordant grains reported by Box et al., (2020) from the upper Deer Trail Group and yields a  $\epsilon\text{Hf}_t$  value of  $+0.2$ .

Detrital zircons grains within all samples show varying levels of lead-loss. Generalized Discordia lines (calculated in IsoplotR; Vermeesch, 2018) for the Gold Cup Quartzite (38DT19), Belt Supergroup (12DTB19), and Deer Trail Group (21JBM19) samples yield lower intercepts from ca. 138 to 78 Ma. These Cretaceous ages likely reflect the timing of burial during the Sevier orogeny, and are generally consistent with peak metamorphism and magmatism within the Priest River Complex (Stevens et al., 2015, 2016).





**Fig. 3.7.** Detrital zircon U–Pb/Lu–Hf results for the Buffalo Hump cobble (34DTB19) shown in a KDE plot and Wetherill concordia diagram. This cobble was collected from the <760 Ma Buffalo Hump Formation and was likely sourced from eroded upper Deer Trail Group (Brennan et al., 2021a). In the Wetherill concordia diagram the green analyses fit within our concordance limits and are reported in the KDE plot. The grey analyses are outside our concordance limits and are omitted from the KDE plot. The concordia inset shows youngest analyses that fit our concordance limits. Solid lines indicated the analyses overlapped with concordia. Representative zircon CL-images and individual analysis results are shown on the right. Approximate average crustal evolution ( $\text{Lu}/\text{Hf} = 0.015$ ) indicated. Abbreviations: DM–depleted mantle, CHUR–chondritic uniform reservoir.

## 3.6. Discussion

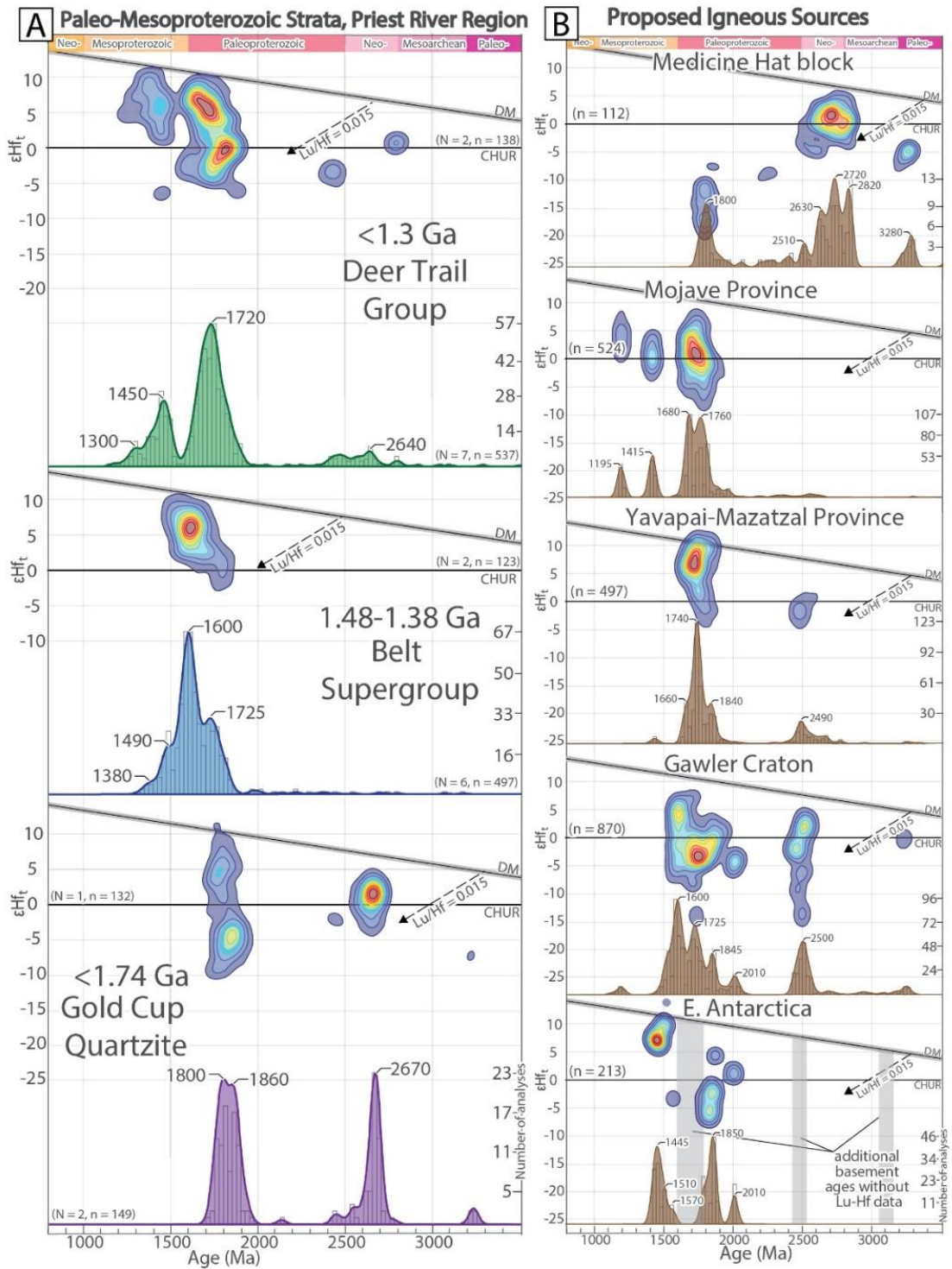
### 3.6.1. The Gold Cup Quartzite: A record of Laurentia assembly prior to final Nuna amalgamation

The relatively high zircon grain aspect ratios within the Gold Cup and Neihart quartzite samples indicate the preservation of elongate grains, suggesting comparably low levels of mechanical abrasion during erosion and transport (e.g. Markwitz and Kirkland, 2018). The detrital zircon spectrum within these samples are dominated by relatively bimodal age distributions at ca. 2.8–2.5 and 1.95–1.7 Ga (Fig. 3.4). These ages are common in the Medicine Hat/Clearwater block, and the Great Falls Tectonic Zone (Fig. 3.8B; Gifford et al., 2020; Vervoort et al., 2016). The Lu–Hf crustal evolution trends suggest sourcing from predominately Meso- and Neoproterozoic crust that was reworked in the late Paleoproterozoic, but which also received broadly coeval juvenile mantle input. The Neihart quartzite sample (82DTB19) contains significantly evolved ca. 1.8 Ga grains with  $\epsilon_{\text{Hf}_t}$  values that range from approximately –10 to –18 (Fig. 9A). Considering that juvenile ca. 1.9–1.6 Ga sources are widespread in Laurentian Trans-Hudson-age arcs and Yavapai–Mazatzal–Mojave terranes in southwestern Laurentia (e.g. Whitmeyer and Karlstrom, 2007), the most evolved

grains within this age range may yield more unique provenance constraints. Many of these most evolved ca. 1.8 Ga grains yield  $\epsilon_{\text{Hf}_t}$  values similar to those in ca. 1.8 Ga rocks from the Medicine Hat Block (Fig. 3.9A; Gifford et al., 2020). Collectively, the zircon grain shape and U–Pb/Lu–Hf analytical results suggest a proximal provenance from the underlying Medicine Hat–Clearwater Block, and the Great Falls Tectonic Zone. Additional inputs from adjacent similar-age sources such as the Wyoming Craton, and poorly characterized (lacking Lu–Hf data) sources such as the Vulcan Structure, Hearne Province (Lovera Block), and the Rimbey Arc are also possible.

Sears et al. (2010) reported a detrital zircon sample ( $n = 95$ ) from the Precambrian quartzite of Argenta, along the southern/southeastern margin of the Belt Basin. This sample also shows a bimodal age distribution, with broad peaks at ca. 2.8 to 2.5 and 2.0–1.7 Ga. Based on these ages Sears et al. (2010) favored a correlation to the Neihart Quartzite, an interpretation supported by our new data. The ca. 1.87 and ca. 2.59 Ga age-peaks in the quartzite of Argenta may reflect input from the Grouse Creek Block (Link et al., 2017; Fig. 3.2A). Comparable zircon ages are present in the Gold Cup and Neihart Quartzite, albeit less prevalent (Fig. 3.10).

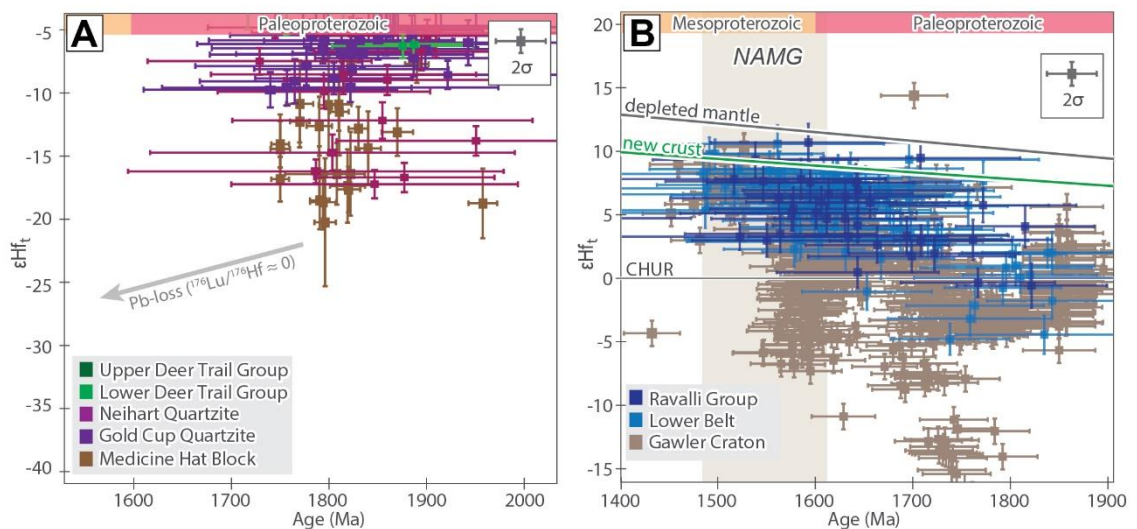
A single sample in which detrital zircons were analyzed ( $n = 22$ ) from a stratigraphic interval of the Fort Steele Formation that (probably conformably) underlies lower Belt Supergroup strata at the northern extent of the Belt Basin also shows ca. 1850 and 2.7–2.6 Ga age-peaks (Fig. 3.10; Ross and Villeneuve, 2003). However, the Fort Steele Formation is significantly thicker and contains finer-grained intervals than the Gold Cup and Neihart Quartzite. Additionally, the Fort Steele Formation contains a single concordant ca. 1635 Ma zircon grain, approximately 100 Myr younger than the youngest grains identified in the Gold Cup, Neihart, and Argenta units.



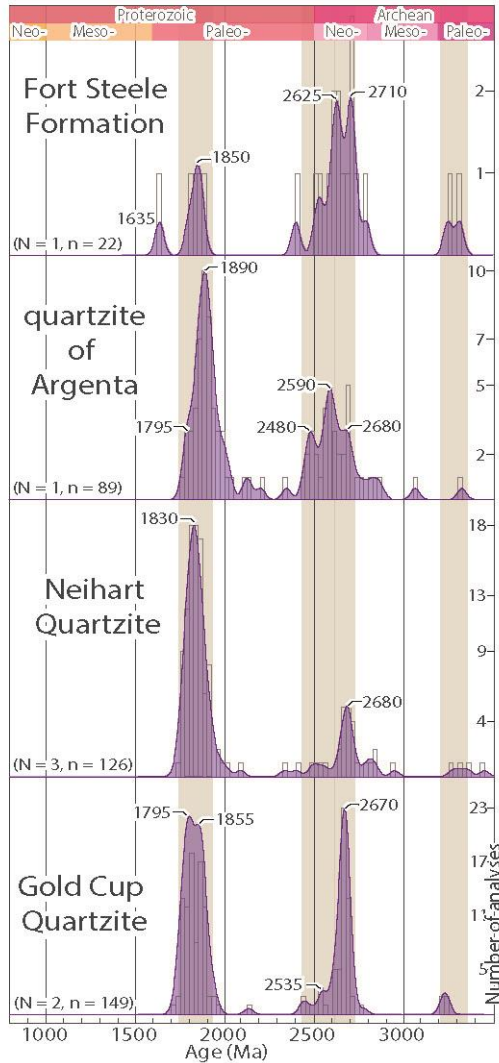
**Fig. 3.8.** A) Detrital zircon U–Pb and Lu–Hf data for Paleo–Mesoproterozoic strata from the Priest River region including our new U–Pb/Lu–Hf data and compiled U–Pb data. Average crustal evolution trends ( $Lu/Hf = 0.015$ ) indicated. Data sources are in Table 1. B) Compiled zircon U–Pb/Lu–Hf data for proposed igneous basement sources. Data source are in the supporting information, any omission of applicable data is non-intentional. Lu–Hf data is plotted using the bivariate kernel density estimate method of Spencer et al. (2020), note that warmer colors (red) indicate greater data density, while cooler colors (blue) indicate lesser data density. Abbreviations: DM–depleted mantle, CHUR–chondritic uniform reservoir, N = number of samples, n = total individual analyses.



Consequently, based on the lithologies and detrital zircon populations of these units, the Gold Cup, Neihart, and Argenta quartzites were likely deposited during a basin-forming event after ca. 1.74 Ga but prior to formation of the main Belt Basin at ca. 1.47 Ga. This period of Neihart–Gold Cup–Argenta sedimentation possibly occurred during final collision and post-collisional collapse of the Great Falls Orogen (e. g. Mueller et al., 2016) and other contemporaneous adjacent collisional zones (such as the Vulcan structure or the Rimbey Arc), and sourced detritus primarily from proximal Laurentian basement. Widespread sedimentation across the Medicine Hat–Clearwater Block and an inferred source from proximal basement rocks and bounding suture zones suggests that the Paleoproterozoic and Archean basement blocks in this region of Laurentia (e.g. Wyoming, Grouse Creek, Medicine Hat, Clearwater, and Hearne) were close to their current configuration by ca. 1.73 Ga. Importantly, in a broader context, these results indicate a widespread Laurentian provenance for the Priest River region between ca. 1.73–1.48 Ga.



**Fig. 3.9.** A) U–Pb and Lu–Hf values for the most evolved ca. 1.8 Ga detrital zircon grains in the Gold Cup (dark purple; 38DTB19) and Neihart Quartzite (light purple; 84DTB19), lower Deer Trail Group (light green; Togo Formation; 01DTB19), upper Deer Trail Group (dark Green; Buffalo Hump cobble; 34DTB19), their likely Medicine Hat Block source (Gifford et al., 2018). B) U–Pb and Lu–Hf values for the ca. 1.9 to 1.45 Ga detrital zircon grains from the Lower Belt (light blue; 10DTB19; Prichard Formation), Ravalli Group (dark blue; 12DTB19; Revett Formation) and compilation of magmatic sources from the Gawler Craton of southeastern Australia (Reid and Payne, 2017; Hartnady et al., 2020).

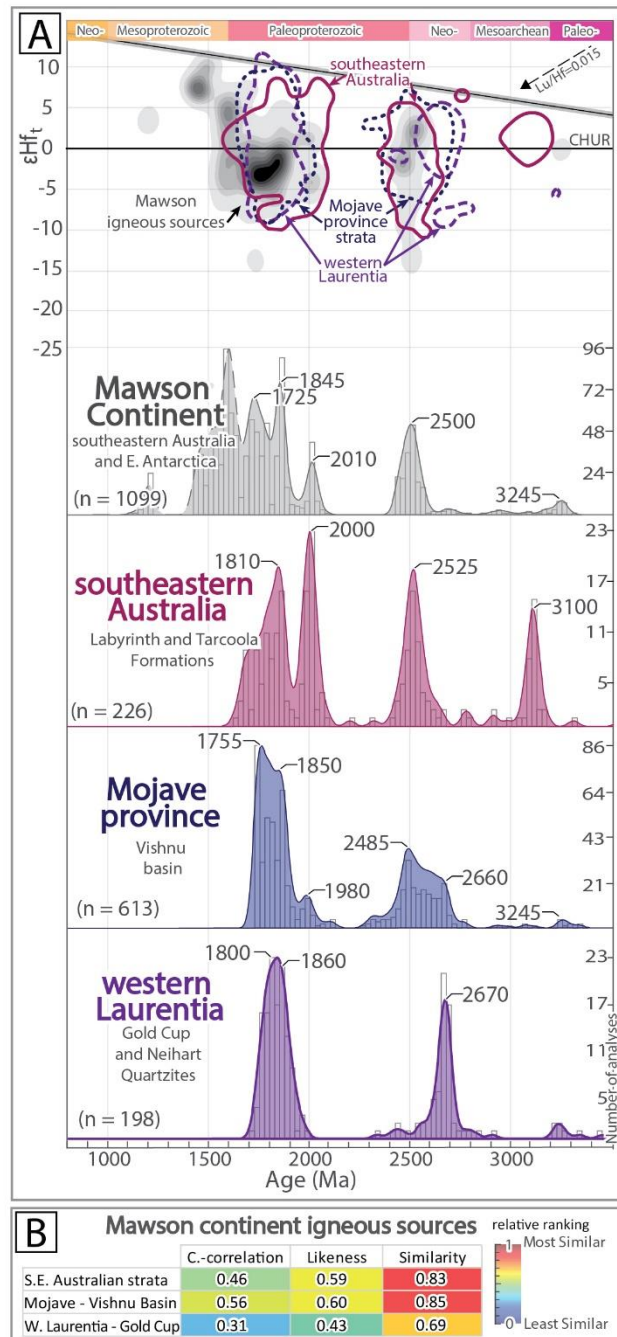


**Fig. 3.10.** Detrital zircon age-spectra for the Fort Steel Formation (Ross and Villeneuve, 2003), Quartzite of Argenta (Sears et al., 2010), Neihart Quartzite (Mueller et al., 2016; this study), and Gold Cup Quartzite (Doughty and Chamberlain, 2008; this study), and their main age-peaks highlighted indicated. These main age-peaks are consistent with proximal sourcing from the Medicine Hat Block and adjacent Paleoproterozoic collisional zones.

However, for this same general time interval, ca. 1.79–1.74 Ga strata in the Mojave province of southwestern Laurentia (e.g. Vishnu Schist), have been suggested to record a non-Laurentian, East Antarctica and Gawler Craton (southeastern Australia) provenance (Holland et al., 2015; 2018). During the Proterozoic, the East Antarctica and Gawler cratons were connected to form a larger continent called Mawson (Cawood and Korsch, 2008; Payne et al., 2009; Fig. 3.1). To evaluate Paleoproterozoic correlation of the Mawson Continent with Laurentia, statistical analysis of the detrital zircon U–Pb/Lu–Hf data from the Vishnu Schist, ca. 1.7 Ga southeastern Australian units, our studied Neihart–Gold Cup units and Mawson (combined East Antarctica–Gawler Craton) igneous sources was conducted. The statistical results (Fig. 3.11B) indicate that Mawson igneous sources show comparable similarity values with the ca. 1.7 Ga Gawler Craton strata and strata of the Vishnu basin in Mojave province. This similarity supports prior correlations of the Mojave province with the Mawson continent (Goodge et al., 2017; Goodge and Fanning, 2016;

Holland et al., 2015, 2018). However, the Gawler Craton igneous sources indicate a much weaker similarity with the Neihart–Gold Cup units. Notably, the presence of the CHUR-like (in regards to Lu–Hf) ca. 2.67 Ga populations within in the Neihart–Gold Cup, that are likely diagnostic of a Laurentian provenance, are not well represented in Mawson rocks.

**Fig. 3.11.** A) U–Pb and Lu–Hf values for (bottom to top): < 1.73 Ga studied western Laurentian strata (Gold Cup, Neihart Quartzite), ca. 1.79 – 1.74 Ga strata in southwestern Laurentia (Vishu basin strata; Holland et al., 2015, 2018; Wooden et al., 2013), ca. 1.7 Ga strata from the Gawler Craton (Howard et al., 2009, 2011; Szpunar et al., 2011), Mawson Continent (southeastern Australia and East Antarctica) igneous sources (Reid and Payne, 2017 and references within). B) Results of two-dimensional (Lu–Hf/U–Pb) statistical comparison (Sundell and Saylor, 2021) of the datasets, with a value of 1 indicating the most similar and 0 the least similar, between Mawson continent igneous sources, and southeastern Australian strata, and southwestern Laurentia (Gold Cup and associated units) and southwestern Laurentia (strata of the Vishu basin). See Sundell and Saylor (2021) for further discussion of the Cross-correlation (C.-correlation), Likeness, and Similarity metrics. Please note that only zircon grains with both U–Pb and Lu–Hf values are included in the compilation and statistical comparison, any omission of applicable data is non-intentional.



Consequently, the U–Pb/Lu–Hf detrital zircon data from ca. 1.79–1.74 Ga Vishnu basin strata of the Mojave province are similar to coeval Australia–Antarctica strata and igneous sources. However, there is no evidence that the Priest River region and the surrounding main Archean blocks (Wyoming, Clearwater, Medicine Hat, Hearne, and Grouse Creek) of western Laurentia were exchanging sediments with Australia–Antarctica during the late Paleoproterozoic. As such, Paleoproterozoic strata in the main Archean blocks of western Laurentia do not support a final assembly of Nuna by ca. 1.8 Ga. Instead, we suggest that the Mojave province had an Australia–Antarctica origin and subsequently joined Laurentia during final Nuna assembly at ca. 1.65–1.6 Ga (Gibson and Champion, 2019). Alternatively, the assembly of Australia–Antarctica with Laurentia may have been a prolonged process that generally propagated from north to south along the western margin of Laurentia (modern coordinates) from ca. 1.75 to 1.6 Ga (c.f. Betts et al., 2011; Furlanetto et al., 2013; Kirscher et al., 2019; Pisarevsky et al., 2014a; Nordsvan et al., 2018).

### **3.6.2. NAMG-age magmatism within the Clearwater Block**

The detrital zircon spectrum of the two analyzed Belt Supergroup samples (Prichard Formation and Ravalli Group) along with existing data (Fig. 3.5; 8A) compiled from the Priest River region, contain a large-proportion of grains that fall within the ca. 1610–1490 Ma NAMG, supporting a western non-Laurentian source. These samples were sourced primarily from juvenile ( $\epsilon\text{Hf}_t = +5$  to  $+7$ ) ca. 1600 Ma and more evolved ( $\epsilon\text{Hf}_t = -5$  to  $+4$ ) ca. 1720 Ma rocks. The only Laurentian sources, the ca. 1580 Ma Laclede Gneiss in the Priest River Complex (Evans and Fischer, 1986), and the recently identified  $1588 \pm 7$  Ma *meta*-rhyolite (the Blue Springs metarhyolite) in the Mazatzal province of southwestern Laurentian (Holland et al., 2020) are localized and do not span the entire age-range present in Belt Supergroup strata. Although there is no Lu–Hf isotopic data available for the Blue Springs metarhyolite, the Laclede Gneiss has a relatively evolved  $\epsilon\text{Hf}_t$  (approximately  $-6$ ; Goodge et al., 2017). Thus, non-Laurentian sources are required to satisfy the range of detrital zircon ages (as previously noted; e.g. Ross and Villeneuve, 2003; Jones et al., 2015; Lewis et al., 2010) and Lu–Hf compositions identified within western Belt Supergroup rocks.

The Gawler Craton of southeastern Australia records widespread magmatism in pulses from ca. 1.9 to 1.5 Ga, which include: 1) the ca. 1850–1790 Ma Donington–Myola Suite (Fanning et al., 1988; Reid et al., 2008), 2) the ca. 1770–1740 Ma McGregor Volcanics and associated suites (Fanning et al., 1988), 3) the 1620–1605 Ma St Peter Suite (Swain et al., 2008), 4) the 1600–1575 Ma Gawler Range Volcanics–Hiltaba suite, and 5) granites as young as ca. 1500 Ma in the Spilsby Suite (Fanning et al., 2007). The ca. 1620–1605 Ma St Peter Suite and ca. 1600–1575 Ma Hiltaba Suite and Gawler Range Volcanics all show distinctly juvenile compositions with average  $\epsilon_{\text{Hf}_t}$  values up to +4 (Fig. 3.8B; Swain et al., 2008; Reid and Payne, 2017).

The juvenile, NAMG-age magmatism of the St. Peter Suite, and Hiltaba and Gawler Range Volcanics are predominately A-type, mantle-derived melts (Chapman et al., 2019) consistent with the U–Pb, and Lu–Hf results for NAMG zircons in the westernmost Belt Supergroup (Fig. 3.9B). Additionally, ( $^{40}\text{Ar}/^{39}\text{Ar}$ ) thermochronology data records widespread exhumation of the Galwer Craton from ca. 1460–1415 (Reid and Forster, 2021), which is the same as the main timing of Belt Supergroup deposition. This strengthens interpretations of a Belt Supergroup–Gawler Craton connection similar to that proposed by Ross et al. (1992), Ross and Villeneuve (2003), Fanning et al. (2009) and Box et al. (2020), within a proto-SWEAT configured Nuna from at least ca. 1.5 to 1.4 Ga (Fig. 3.12B).

### **3.6.3. Provenance and age of the Deer Trail Group: A ca. 1.3–1.2 Ga sequence deposited after Nuna breakup but likely prior to Grenville-age tectonism**

Upper Belt Supergroup strata (Missoula and Lemhi groups) are intruded by ca. 1370 Ma granites in east–central Idaho and lack detrital zircon grains younger than ca. 1380 Ma (Doughty and Chamberlain 2008; Stewart et al., 2010; Link et al., 2016). Thus, the  $1299 \pm 12$  Ma maximum depositional age for the basal Deer Trail Group falsifies initial correlations of the Deer Trail Group to Piegan and lower Missoula group strata of the Belt Supergroup (Miller and Whipple, 1989). Furthermore, the ca. 80 Myr gap between depositional age constraints on the ca. 1.38 Ga upper Belt Supergroup rocks and the younger basal Deer Trail Group is similar in age and duration to the enigmatic East Kootenay tectono-thermal event. This may suggest that the East Kootenay event separated (or spanned) these two basin-forming intervals. Grenvillian (ca. 1.2–1.0 Ga) detrital zircons are common in nearby Neoproterozoic

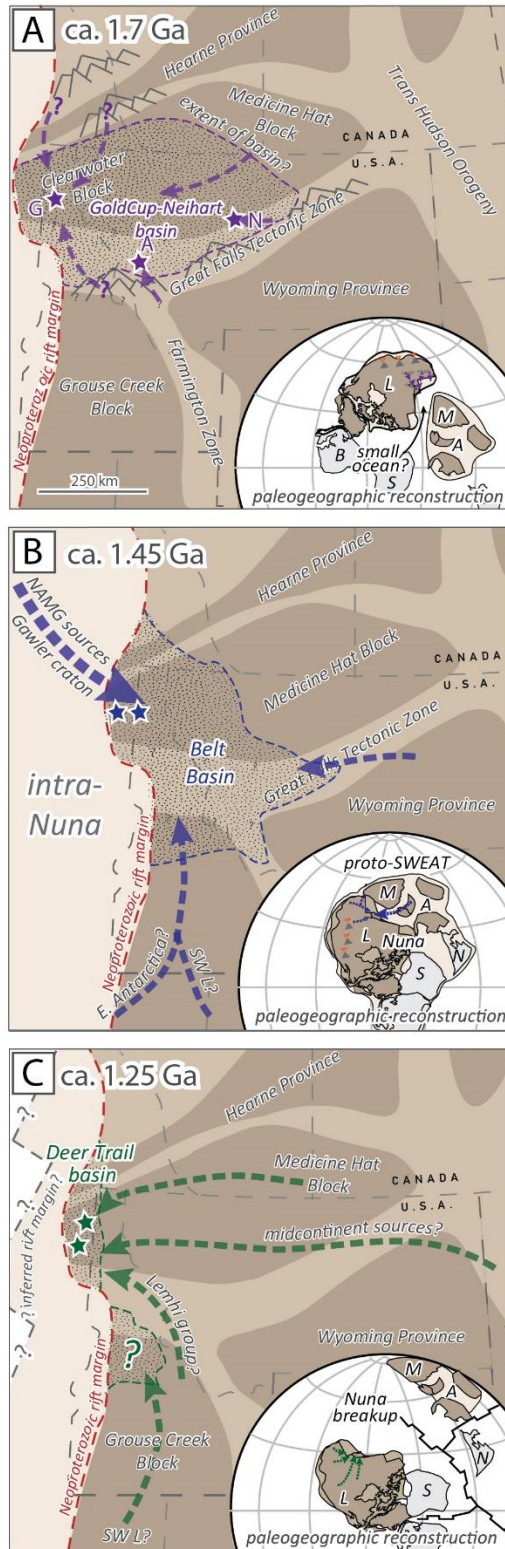
strata (e.g. Yonkee et al., 2014; Matthews et al., 2018; Box et al., 2020; Brennan et al., 2020; 2021a; 2021b) but are absent in the Deer Trail Group. Consequently, we suggest that deposition of the Deer Trail Group likely occurred prior to the ca. 1.1 Ga main collisional phase of Grenville orogeny in eastern/southeastern Laurentia (e.g. Tollo et al., 2004; Mulder et al., 2017). Alternatively, the Deer Trail Group could be a younger (<1.1 to 0.76 Ga) basin that was not fed by a Grenvillian-source but this is less likely given the relative abundance of Grenville-age detrital zircon grains in strata of this age.

The overall fine-grained siliciclastic and interbedded carbonate character of the Deer Trail Group stratigraphy is consistent with deposition in a shallow intracratonic basin or marine continental shelf depositional environment (Box et al., 2020). The prevalence of soft-sediment deformation may also signify slumping, perhaps influenced by seismically-induced slope instability (e.g. Bose et al., 2012; Pratt and Ponce, 2019; Pratt and Rule, 2021). The broad ca. 1.9–1.6 Ga and 1.45 Ga detrital zircon populations in the Deer Trail Group are similar in age to those found in the Missoula Group (and the correlative Lemhi Group) of the Belt Basin. However, the range of Lu–Hf values of the ca. 1.9–1.6 Ga detrital zircon grains within the Deer Trail Group are a poor match for any single basement source (Fig. 3.8). Consequently, to satisfy the range of ages and Lu–Hf values of the ca. 1.9 to 1.6 Ga zircon grains from Deer Trail Group strata, a mixed provenance from the Medicine Hat Block, Mojave province, and/or recycling from ca. 1.7 Ga Gold Cup/Neihart or ca. 1.4 Ga Lemhi Group strata is required (Figs. 3.8 and 3.12C).

The ca. 1.5–1.3 Ga grains within the Deer Trail Group samples contain  $\epsilon\text{Hf}_t$  values mostly between 0 to +12, with a few grains containing evolved ( $-\epsilon\text{Hf}_t$ ) values. The ca. 1.5–1.3 Ga Laurentian midcontinent granite–rhyolite province in southwestern Laurentia contains sources of similar ages with mostly juvenile  $\epsilon\text{Hf}_t$  values, consistent with crustal melts sourced from the late Paleoproterozoic juvenile Yavapai–Mazatzal–Mojave provinces they intrude (Bickford et al., 2015). Generally, the most evolved ca. 1.45 Ga midcontinent granites intrude along the southeastern edge of the Superior province, in the midcontinent region (Bickford et al., 2015). However, the few detrital zircon grains with particularly evolved  $\epsilon\text{Hf}_t$  values (below  $-5$ ) within the Deer Trail Group, despite still being within our discordance filter, show a higher level of discordance (generally  $>5\%$  discordant), suggesting these grains may not provide robust provenance constraints. Consequently, the Deer Trail Group detrital zircon



components may indicate a mixed distal Laurentian and proximal recycled provenance. Notably, *syn*-depositional magmatism is limited or absent within the Deer Trail Group, and the fine-grained nature of these strata likely precludes significant proximal uplift.



**Fig. 3.12.** Proposed sedimentary transfer systems (indicated by arrows) to the Priest River region during the A) early assembly, B) tenure, and C) breakup of Nuna with global paleogeographic reconstruction cartoons adopted from Kirscher et al. (2020). At ca. 1.7 Ga (A) during deposition of the Gold Cup Quartzite, and likely coeval Neihart Quartzite, Quartzite of Argenta, and perhaps Fort Steele Formation, proximal Paleoproterozoic suture zones (Great Falls Tectonic Zone; Vulcan Zone, Rimbey Arc) likely supplied sediment into a basin(s) that likely extended across most of the Medicine Hat–Clearwater Block. Abbreviations: A–Quartzite of Argenta, N–Neihart Quartzite, G–Gold Cup Quartzite, S–Fort Steele Formation. (B) At ca. 1.45 Ga during deposition of Belt Supergroup strata. Belt Supergroup strata in the Priest River region are characterized by NAMG detrital zircon populations consistent with sources from Australia–Antarctica, notably the Gawler craton. Belt strata along the eastern and southern extent of the basin, generally lack NAMG zircon grains, and contain ages reflecting southwestern Laurentia (SWL; e.g. Ross and Villeneuve, 2003; Ronemus et al., 2020) or perhaps E. Antarctica sources (Stewart et al., 2010; Link et al., 2016). (C) At ca. 1.3 Ga during deposition of Deer Trail Group strata. These rocks generally indicate a mixed Laurentian provenance, and are consistent with depositional during the final interval of the ca. 1.38–1.30 Ga East Kootenay tectono-thermal event. We propose the East Kootenay orogeny recognized in this region, is consistent with Doughty and Chamberlain’s (2008) model of renewed rifting, and subsidence in the Priest River area. This event likely resulted in the end of Belt Supergroup deposition and shortly preceded deposition of the Deer Trail Group coeval with the proposed timing of widespread Nuna breakup (Pisarevsky et al., 2014a,b; Kirscher et al., 2020).

At first order, these stratigraphic observations are less consistent with collisional models for the ca. 1.38–1.30 Ga East Kootenay “orogeny” (McFarlane and Corfu, 2015; McMechan and Price, 1982). Instead, these stratigraphic constraints are more consistent with the East Kootenay “orogeny” as a period of basin rifting and renewed subsidence that marked the end of deposition within the Belt Basin (Doughty and Chamberlain, 2008). We suggest this event shortly preceded deposition of the Deer Trail Group coeval and was coeval with the proposed timing of widespread Nuna breakup at ca. 1.3 Ga (Pisarevsky et al., 2014b; Kirscher et al., 2020). If correct, this implies a late-rift to passive margin setting for deposition of the Deer Trail Group. However, given the relatively limited minimum depositional age-constraints, this is currently speculative. It is unclear how extensive the Deer Trail sequence may have been, although vestiges of its original extent may be present in central Idaho (Isakson, 2017; the “?” basin in Fig. 3.12C).

#### **3.6.4. Smaller block within “proto-SWEAT” reconstructions**

Various smaller crustal blocks, notably Hainan Island (with or without the entire Cathaysia Block; Li et al., 1995; 2008a; b; Xu et al., 2019; Yao et al., 2017) and Tasmania (Halpin et al., 2014; Mulder et al., 2015) are suggested to have occupied the region between western Laurentia and East Antarctica/Australia in a proto-SWEAT configured Nuna. The “Missing-Link” model proposes that the Cathaysia Block of South China was an extension of western Laurentia from ca. 1.8 Ga until Rodinia breakup (Li et al., 1995; 2008a). In particular, this model suggests that ca. 1.4 Ga juvenile granites on Hainan Island (Li et al., 2008b; Yao et al., 2017) represent a continuation of the A-type granitic province of southwestern Laurentia (e.g. Bickford et al., 2015). Additionally, the Baoban Complex metasedimentary rocks (including the generally coeval Shilu Group) that also contain NAMG-age detrital zircon age-components (Xu et al., 2019; Yao et al., 2017) are suggested correlatives to Belt Supergroup strata of western Laurentia. However, the Proterozoic configuration of South China, including whether Hainan Island was separate from or part of the Cathaysia or proto-Yangtze Block prior to the mid-Paleozoic is debated (c.f. Yao et al., 2017; Cawood et al., 2020; Liu et al., 2020; Xu et al., 2019). Consequently, Hainan Island may not be a reliable constraint for the location of the larger South China blocks.

On Hainan Island, the ca. 1.4 Ga paragneiss (the Gezhencun “succession”) of the Baoban complex contains ca. 1.9–1.4 Ga detrital zircon grains, including some ca. 1.6–1.5 Ga NAMG grains. Based on limited Lu–Hf data ( $n = 21$ ; Xu et al., 2019), these



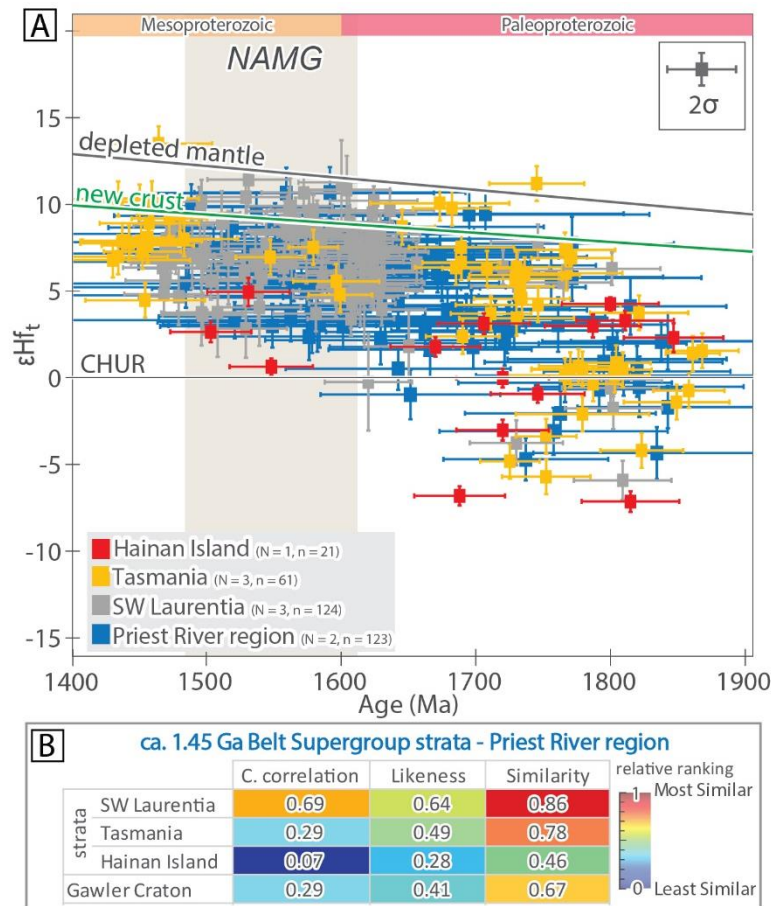
ca. 1.9–1.5 Ga detrital zircon grains show  $\epsilon\text{Hf}_t$  values less than +5 (Fig. 3.13A). Consequently, the juvenile ( $\epsilon\text{Hf}_t = +5$  or greater) ca. 1.9 to 1.5 Ga grains that comprise approximately 50% of the total detrital zircon cargo in Belt Supergroup strata within the Priest River region are not recognized in coeval rocks on Hainan Island (Yao et al., 2017).

The identification of NAMG-age detrital zircon components in the lower–middle Rocky Cape Group of Tasmania led Halpin et al. (2014) and Mulder et al. (2015) to correlate the lower–middle Rocky Cape Group of Tasmania to the Belt Supergroup and Deer Trail Group of Laurentia. This interpretation has Tasmania located in a central position between western Laurentia and East Antarctica/southeastern Australia within Nuna (Halpin et al., 2014; Mulder et al., 2015). The juvenile ( $\epsilon\text{Hf}_t = +5$  or greater) ca. 1.8 to 1.5 Ga grains that characterize Belt Supergroup strata in the Priest River region, and coeval strata in southwestern Laurentia (Doe et al., 2013) are present within the lower Rocky Cape Group (Fig. 3.13; Mulder et al., 2015), consistent with similar proposed Gawler Craton–East Antarctica (Mawson continent) sources. Consequently, Tasmania’s proposed location adjacent to southwestern/western Laurentia (e.g. Halpin et al., 2014; Mulder et al., 2015; Fig. 3.1) at ca. 1.5 Ga is consistent with existing provenance constraints.

Statistical analysis of the detrital zircon (U–Pb/Lu–Hf) data from these ca. 1.45 Ga basins in Laurentia, Tasmania, Hainan Island and their proposed Gawler Craton source (Fig. 3.13B) allows objective evaluation of these proposed correlations. Overall, the statistical results indicate that the similarity of detrital zircon components in the Belt Basin and coeval southwestern Laurentian basins (Trampas, Yankee Joe–Blackjack basins) is the strongest as they consistently yield the greatest (closest to 1) comparison values (Fig. 3.13B). Thus, it is probable that the ca. 1.45 Ga Yankee–Joe Defiance Basin (Doe et al., 2013) in southwestern Laurentia and the Belt Basin, ~1500 km to the north, received sediment from a similar (non-Laurentian) source. This source was likely the Gawler Craton of southeastern Australia, and perhaps associated rocks in East Antarctica (Jones et al., 2015).

The next strongest statistical resemblance to the Belt Supergroup detrital zircon dataset is coeval strata of the lower Rocky Cape Group in Tasmania (Mulder et al., 2015). Gawler Craton igneous sources also generally show statistical similarity values (to the Belt Supergroup detrital zircon dataset) that are comparable to the values given by the Tasmanian strata (Fig. 3.13B). Consequently, the new data from the Belt

Supergroup is consistent with prior interpretations that Tasmania occupied a central position between western Laurentia and East Antarctica/southeastern Australia within Nuna (Halpin et al., 2014; Mulder et al., 2015). Based on the relatively limited detrital zircon Lu/Hf available from the Gezhencun succession of Hainan Island (Xu et al., 2019), the weakest correlation is between Belt Supergroup strata and coeval strata on Hainan Island. This weak statistical correlation likely reflects to the absence of juvenile ( $\epsilon\text{Hf}_t = +5$  or greater) ca. 1.6–1.5 Ga detrital zircon grains in the Hainan Island strata which are present in the Laurentian and Tasmanian strata, and the Gawler Craton igneous sources. Consequently, prior assertions of the correlation between Belt Supergroup rocks and coeval rocks on Hainan Island (e.g. Xu et al., 2019; Yao et al., 2017) are the least justified.



**Fig. 3.13.** A) Detrital zircon U–Pb and Lu–Hf data from ca. 1.45 Ga strata from Hainan Island, South China (the Gezhencun succession, Xu et al., 2019), Western Tasmania (lower Rocky Cape Group; Mulder et al., 2015), southwestern Laurentia (the Blackjack Formation and Defiance Quartzite; Doe et al., 2013), and Belt Supergroup strata of the Priest River region (Prichard and Revett formations). B) Results of two-dimensional (Lu–Hf/U–Pb) statistical comparison (Sundell and Saylor, 2021) of the datasets above, with a value of 1 indicating the most similar and 0 indicating the least similar between Belt Supergroup strata of the Priest River region, and strata of southwestern Laurentia, Tasmania, Hainan Island, and igneous sources within the Gawler Craton. See Sundell and Saylor (2021) for further discussion of the Cross-correlation (C.-correlation), Likeness, and Similarity metrics.

### 3.7. Conclusions

U–Pb and Lu–Hf detrital zircon provenance data from the (<1.73 Ga) Gold Cup Quartzite, (ca. 1.47–1.38 Ga) Belt Supergroup and overlying (<1.3 Ga) Deer Trail Group of the Priest River region (northern Idaho, and northeastern Washington state, USA) give insights into the regional tectonostratigraphy of western Laurentia, and yield important constraints on the assembly of western Laurentia and the evolution of supercontinent Nuna. Our major findings are the following:

1) The Gold Cup Quartzite, Neihart Quartzite, and the quartzite of Argenta show similar detrital zircon age spectra that suggests sourcing from proximal Neoproterozoic and Paleoproterozoic Laurentian basement. Such similarities, and the general absence of zircon grains younger than ca. 1.73 Ga within these units, suggest that assembly of the main Archean–Paleoproterozoic blocks of western Laurentia (Wyoming, Clearwater, Medicine Hat, Hearne, and Grouse Creek) was likely complete by ca. 1.73 Ga.

2) The Laurentian < 1.73–1.48 Ga, Gold Cup and Neihart quartzites do not have require non-Laurentian sources. Thereby, this data does not compel a proto-SWEAT configured Nuna by ca. 1.8 Ga as previously suggested (e.g. Rogers and Santosh, 2002; Zhao et al., 2002; Holland et al., 2018), and instead favors a protracted, perhaps two-stage assembly of Nuna from ca. 1.8 to 1.6 Ga (e.g. Kirscher et al., 2019).

3) Ca. 1.47 Ga strata in the western extent of the Belt Basin contain juvenile ca. 1610–1490 Ma “North American magmatic gap” zircon grains that lack a Laurentian source. The Gawler Craton contains igneous sources that span the required age and Lu–Hf compositions of these grains, and experienced active exhumation at ca. 1.47 Ga. Consequently, our new constraints are consistent with suggestions that southeastern Australia was a source of sediment for Belt Supergroup strata. This data supports proto-SWEAT reconstructions for Nuna that place the Gawler Craton of southeastern Australia adjacent to the Priest River region of Laurentia by at least ca. 1.47 Ga.

4) Deposition of the Deer Trail Group occurred after ca. 1.3 Ga, perhaps coeval with the end or latter part of the enigmatic East Kootenay tectono-magmatic event. The ages and range of Lu–Hf values recorded by the main detrital zircon components within the Deer Trail Group suggest a mixed provenance from several Laurentian sources, including the Mojave Province, Trans-Hudson Orogen, and perhaps the

Medicine Hat Block. Many of these grains could have been recycled from nearby older strata, such as the Lemhi Group of the Belt Basin. The fine-grained siliciclastic and carbonate lithology and the lack of significant *syn*-depositional magmatic detrital zircon components in Deer Trail Group strata are consistent with their deposition during a period of regional subsidence within the Priest River region that coincides with purported Nuna breakup at ca. 1.3 Ga (Kirscher et al., 2020).

Collectively, these new provenance constraints are most consistent with paleogeographic models that advocate widespread assembly of the main Archean blocks of western Laurentia from ca. 1.9 to 1.7 Ga, and indicate that the Priest River region does not record sedimentation from non-Laurentian source terranes until after ca. 1.5 Ga. Consequently, the Paleo–Mesoproterozoic tectonostratigraphic record of the Priest River region is consistent with models that indicate the main assembly of western Laurentia was a precursor to, and not necessarily synchronous with, the final assembly of a proto-SWEAT configured Nuna supercontinent.

### 3.8. References

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## **Chapter 4: Grenville-age metamorphism within the Belt Basin of western Laurentia**

### **Abstract**

The presence or absence of ca. 1100–1000 Ma “Grenville-age” tectonism in western Laurentia is long debated and has significant implications for the assembly of supercontinent Rodinia. Recently, identification of ca. 1350–1020 Ma garnet-grade metamorphism within the Clearwater complex of the Belt Basin has renewed this debate. However, the tectonic implication of such cryptic metamorphic records is uncertain due to regional structural and thermal overprinting by younger Cordilleran processes. Here, we present new monazite and apatite U–Pb results from metapelites within the western region of the Clearwater complex. Although monazite ages overlap with existing garnet ages, some apatite records ca. 680 Ma ages associated with nearby rift-related magmatic rocks that formed during Rodinia breakup and indicate that locally Cordilleran conditions did not exceed ~450–550°C in the western Clearwater complex. Consequently, the mineral assemblages within these rocks most likely reflect the peak metamorphic event recorded by the youngest ca. 1100–1020 Ma garnet Lu–Hf and monazite U–Pb ages. Thermodynamic modeling of these assemblages indicates P/T conditions of 5–7.5 Kbar and 620–670°C, suggesting burial depths of at least ~17–22 km along a moderately-high average geothermal gradient of ~23–36°C/km. These ca. 1100–1020 Ma metamorphic conditions are difficult to satisfy with thermal upwelling/heating within the Belt Basin alone, and thus advocate for western tectonism and localized Grenville-age structural inversion and crustal thickening within the Belt Basin, likely along inherited structures. Consequently, global geologic models should account for ca. 1100–1020 Ma Grenville-age tectonism in continental blocks adjacent to the western margin of Laurentia within supercontinent Rodinia.

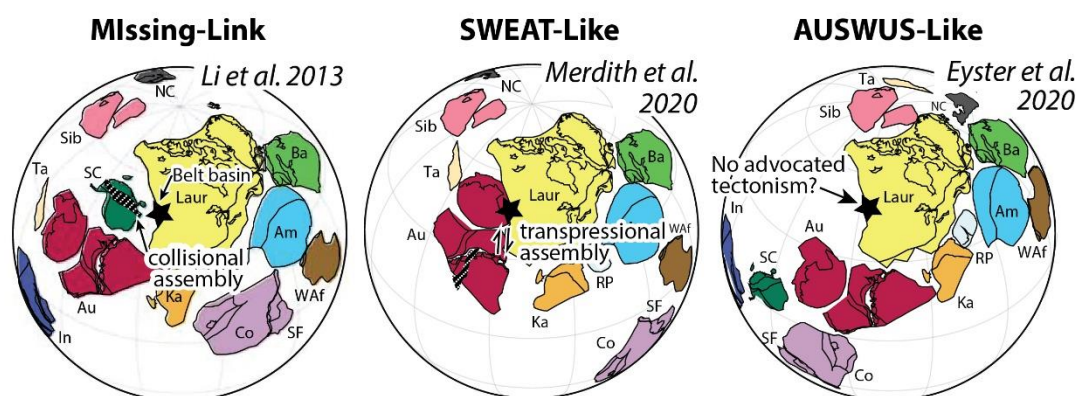
## 4.1. Introduction

The Mesoproterozoic Era (1600–1000 Ma) records the transition from Earth's first supercontinent Nuna (also called Columbia; ca. 1600–1300 Ma) to its successor, Rodinia (ca. 1100–800 Ma; e.g. Li et al., 2019). However, researchers have suggested that the global transition from Nuna to Rodinia may have significant geodynamic differences from the previous assembly of Nuna, and the ensuing Rodinia to Pangea supercontinent transition. Notably, the geologic record suggests evidence for a different style of plate tectonics associated with systemically higher thermobaric (temperature/pressure; T/P) conditions during metamorphism (Brown et al., 2020), a warmer upper mantle and the presence of hot, low and thin “Grenville-age” mountain belts (Spencer et al., 2021; Zhu et al., 2022), and/or the closure of “young” interior oceans, rather than older exterior oceans during this supercontinent transition (e.g. Li et al., 2019). Consequently, the Nuna to Rodinia supercontinent cycle is considered to be a perplexing geological time interval dubbed the “Boring Billion” (Cawood and Hawkesworth, 2014; Sobolev and Brown, 2019).

The ca. 1480–1380 Ma Belt Supergroup outcrops along the western margin of Laurentia and represents one of the largest Mesoproterozoic sedimentary sequences in the world, containing an immense thickness of strata (>15 km), covering >200,000 km<sup>2</sup> (Lonn et al., 2020). The monotonous lithology of Belt Supergroup stratigraphy records its deposition in an intracratonic rift basin near the center of the Nuna supercontinent landmass (Winston and Link, 1993; Box et al., 2020). Enigmatically, after deposition, despite paleogeographic models calling for globally widespread tectonism associated with the subsequent breakup of Nuna and assembly of Rodinia, most Belt Supergroup strata remained largely undisturbed for the next 1 billion years until the late Mesozoic Cordilleran (Sevier/Laramide) orogeny. Consequently, the general ca. 1300–900 Ma tectonic quiescence in this region, notably the absence of ca. 1100–1000 Ma Grenville-age tectonism, may indeed lend credence to the assertion that the Nuna to Rodinia transition was “boring”, at least in this region.

Then again, isotopic studies along the western portion of the Belt Basin, particularly within strata recently buried deep within Cordilleran metamorphic core complexes (notably the Matthew Creek metamorphic zone and associated Priest River metamorphic complex, the Clearwater metamorphic core complex and the Bitterroot metamorphic core complex) have revealed a growing body of evidence for tectonism

within this interval. This evidence includes indications of ca. 1115–1065 Ma Grenville-age metamorphism (e.g. Doughty and Chamberlain, 2008; Zirakparvar et al., 2010; Nesheim et al., 2012; Slack et al., 2020) that has notable implication for conflicting collisional (e.g. Li et al., 1995; 2013; Yao et al., 2017), transpressional (e.g. Merdith et al., 2020; Mulder et al., 2018) and/or apparently tectonically quiescent (e.g. Eyster et al., 2018) models of Rodinia assembly along western Laurentia (Fig. 4.1). However, interpretations of the importance of the Grenville-age isotopic ages within the Belt Basin, and consequently evaluation of proposed global geodynamic models are often in disagreement. Reaching a consensus on their importance is likely hampered by difficulties in separating the younger, Cordilleran record of metamorphism from the older, Grenville-age record. Some workers interpret the Grenville-age metamorphism to simply reflect static magmatic heating at depth (e.g. Doughty and Chamberlain, 2008; Slack et al., 2020), while others have suggested that the ages require tectonic thickening (Zirakparvar et al., 2010; Neishem et al., 2012).



**Fig. 4.1:** Leading models with differing for Rodinia configurations and assembly mechanism. These models include: Missing-Link collisional (Li et al., 1995; 2013; Yao et al., 2017), SWEAT-like (Southwest U.S.–East Antarctica; Moores, 1991) transpressional (e.g. Mulder et al., 2018; Merdith et al., 2020) and/or apparently AUSWUS-like (Australia–Southwestern U.S.; Karlstrom et al., 1999) tectonically quiescent (e.g. Eyster et al., 2018) mechanism of Rodinia assembly along western Laurentia. Abbreviations include: Am, Amazonia; Au, Australian cratons including Antarctic Mawsonland; Ba, Baltica; Co, Congo; In, India; Ka, Kalahari; Laur, Laurentia; NC, North China; RP, Rio Plata; SC, South China; SF, Sao Francisco; Sib, Siberia; Ta, Tarim; Waf, West African craton.

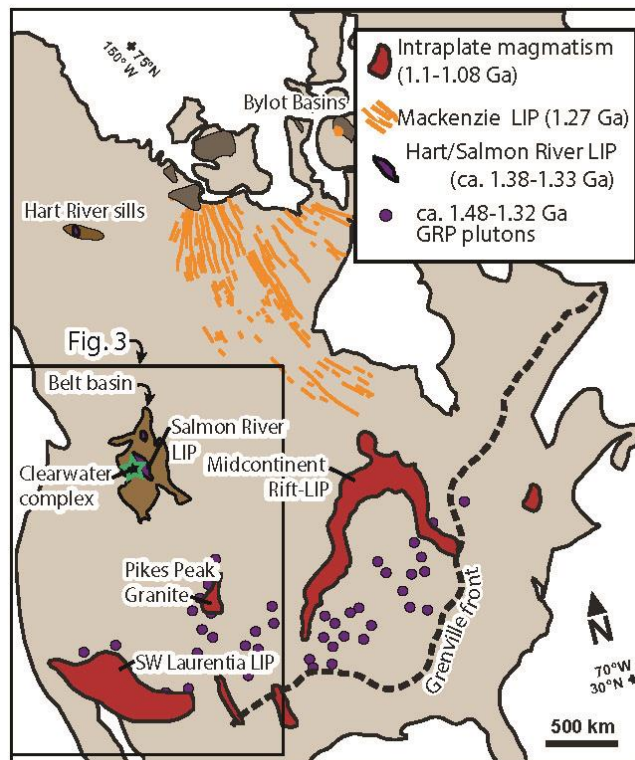
Here, we present new laser-ablation inductively coupled mass spectroscopy (LA–ICPMS) monazite and apatite U–Pb geochronology, electron probe microanalysis (EPMA) of mineral compositions and phase equilibrium P–T modeling of metapelitic strata from the Clearwater complex. These rocks contain a record of ca. 1350–1000 Ma garnet growth under unknown conditions (Zirakparvar et al., 2010; Neishem et al., 2012), which is variably overprinted by younger ca. 130–55 Ma

Cordilleran metamorphism that reaches as high as lower granulite facies (Doughty et al., 2007). Collectively, our results were able to constrain a portion of the complex that was not buried deeply within the Cordilleran orogeny and thus preserves a critical record of amphibolite-grade Grenville-age metamorphism in this region. These results reveal important insights into the Nuna to Rodinia transition in western Laurentia, and provide additional constraints on evaluating global tectonic and metamorphic processes during the late Mesoproterozoic.

## **4.2. Geologic Overview**

The North American Cordillera is a sinuous mountain belt that runs for ~6500 km along the western edge of Laurentia, from Alaska to Mexico. The orogen records prolonged subduction of oceanic lithosphere beneath western North America since the early Mesozoic. Along most of the Cordillera, Late Cretaceous to Paleogene overthickening of the orogenic wedge and/or changing subduction boundary conditions (e.g. flat slab subduction) resulted in the formation of metamorphic core complexes (Dickinson, 2004; Fig. 4.3). These core complexes expose rocks formerly buried deep within the hinterland of the Cordilleran orogen due to late Mesozoic or younger crustal thickening (e.g. Wernicke, 1981; Stevens et al., 2015; Howlett et al., 2021). Although the Cordillera is an archetypal natural laboratory to study modern orogenic processes, overprinting by Cordilleran processes commonly obscures the record of older events along the western margin of Laurentia. Unfortunately, the western margin of Laurentia occupies a critical, central, position along which most models advocate the final assembly of Proterozoic supercontinents Nuna and Rodinia occurred. As a result, difficulty in deciphering the regional pre-Cordilleran record of western Laurentia has made evaluation of conflicting global models challenging.

Following a tectonic lull during the ca. 1610–1490 Ma North American Magmatic Gap associated with final Nuna assembly (Ross and Villenavue, 2001; Brennan et al., 2022), juvenile terranes and arcs were accreted along most of the southern margin of Laurentia from ca. 1480–1350 Ma. Accretion was associated with widespread bimodal ‘A-type’ granite and anorthosite magmatism within the Granite-Rhyolite province (e.g. Bickford et al., 2015; Fig. 4.2). In southwestern Laurentia, magmatism was contemporaneous with the ca. 1460–1400 Ma Picuris orogeny (Karlstrom et al., 2001; Daniel et al., 2013) then, from ca. 1380–1320 Ma, development of an extensive backarc system (Mulder et al., 2017).



**Fig. 4.2:** Simplified geologic map of Laurentia (adapted from Greenman et al., 2021) indicating the relevant ca. 1480–1080 Ma geologic provinces including the ca. 1480–1380 Ma Belt basin, ca. 1480–1320 Ma Granite Rhyolite province, ca. 1380–1330 Ma Hart and Salmon River Large Igneous Provinces (LIP), ca. 1270 Ma Mackenzie LIP and the ca. 1100–1080 Ma Midcontinent Rift/LIP, Pikes Peak Batholith, Southwestern Laurentian LIP (SWLLIP), and Arctic Bylot Basins. The extent of the North American Cordillera shown in Fig. 4.3 is indicated.

The ca. 1470–1380 Ma Belt Basin is an extensive intracratonic basin that formed within the supercontinent Nuna, concurrent with the aforementioned more widespread tectonomagmatic activity in southern Laurentia, and contains over 15 km thick of shallow water and subaerially deposited strata, that record intermittent, localized syndepositional magmatism (Lonn et al., 2020 and references within). Following Belt Basin deposition, continental-arc magmatism occurred in southern Laurentia from ca. 1250–1230 Ma due to continued subduction beneath the southern margin of Laurentia that predated onset of the main episode of Grenvillian orogenesis.

Grenville-aged tectonism in southern Laurentia peaked at ca. 1140–1100 Ma (Mulder et al., 2017), broadly coeval with slab break-off and/or delamination beneath southern Laurentia (Mosher et al., 2008). Subsequent impingement of a mantle plume beneath the Laurentian midcontinent resulted in formation of the ~3000-km long Midcontinent Rift (Cannon and Hinze, 1992; Stein et al., 2015) and emplacement of ca. 1094–1080 Ma diabase dykes of the Southwestern Laurentian large igneous province (SWLLIP) (Bright et al., 2014). The record of magmatism at ca. 1080 Ma extends as far northwest as central Colorado, where the ca. 1115–1065 Ma Pikes Peak batholith intruded along crustal sutures within the Yavapai province, and likely records lithospheric delamination and/or plume activity (Guitreau et al., 2016). In northern Laurentia, intracratonic basins (the Bylot basins) developed from ca. 1090 Ma to 1050

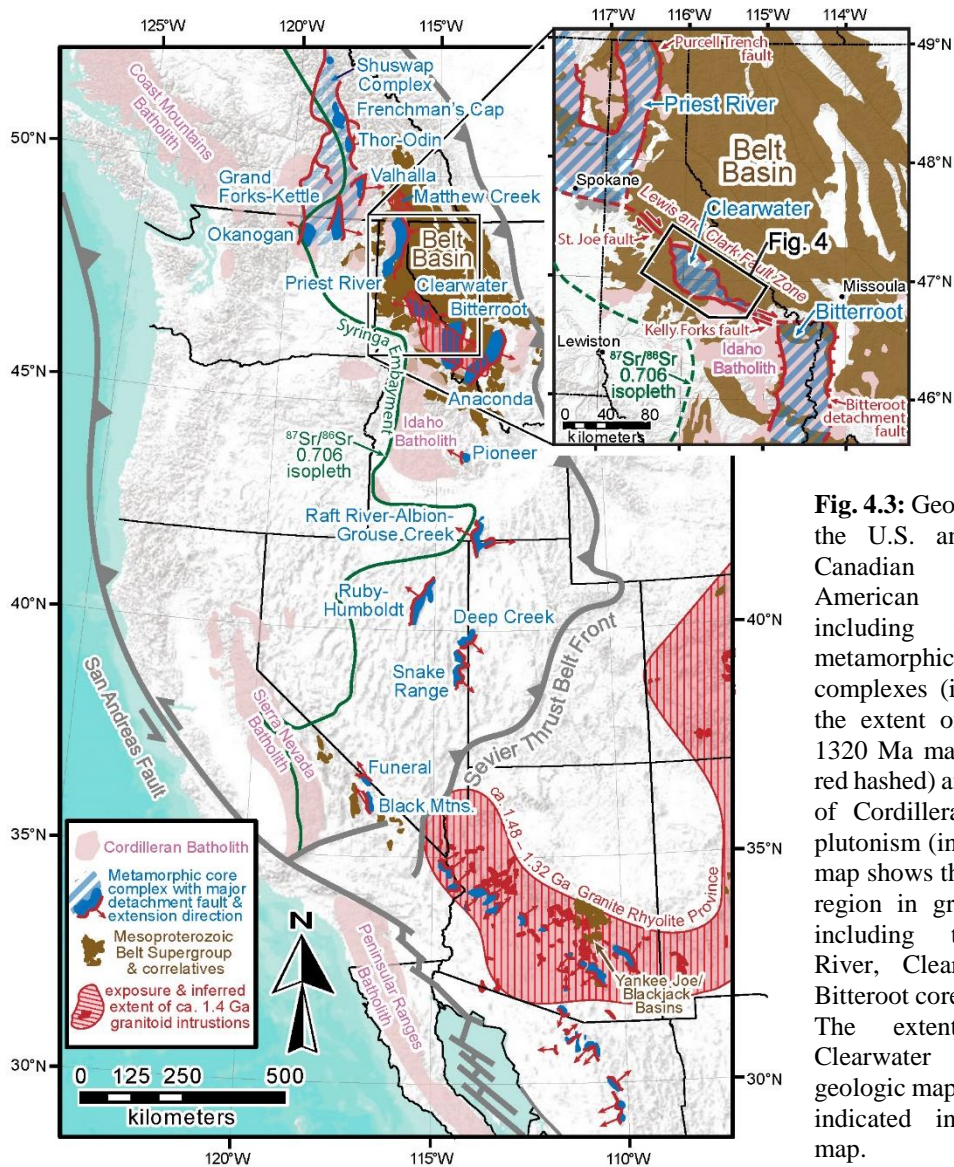
Ma (Fig. 4.2; Gibson et al., 2019; Rainbird et al., 2020; Greenman et al., 2021). Collectively, the data are consistent with a widespread episode of thermal upwelling and localized extension across Laurentia around ca. 1120–1050 Ma, in a period otherwise dominated by Grenvillian orogenesis associated with assembly of Rodinia. Grenville-age (ca. 1300–1000 Ma) orogens are recognized on many continental blocks (such as Laurentia, South China, Australia, Kalahari and Baltica) and have long been known to reflect the widespread assembly of supercontinent Rodinia (e.g. Moores, 1991; Li et al., 2008; Mulder et al., 2017; Johansson et al., 2022).

The Clearwater complex is the central of three main core complexes that occur along the western edge of the Belt Basin, including the Priest River complex and associated Matthew Creek metamorphic zone ~150 km to the north, and the Bitterroot complex ~150 km to the southeast (Fig. 4.3). These complexes all record an enigmatic record of ca. 1400–1000 Ma tectonism, which is variably overprinted by younger Cordilleran processes, but is generally attributed to two main events. These two events consist of an older ca. 1380–1300 Ma East Kootenay “Orogeny” (McMechan and Price, 1982) and a younger poorly understood ca. 1115–1060 Ma Grenville-age event (e.g. Neishem et al., 2012).

The ca. 1380–1300 Ma East Kootenay tectono-magmatic event was contemporaneous with the final stages of tectonism and magmatism within the Granite-Rhyolite province in southern Laurentia, and largely records high-T, low-P regional metamorphism (McFarlane and Pattison, 2000; McFarlane, 2015) and bimodal magmatism within the Salmon River sills and associated Elk City Domain within the Belt Basin (Doughty and Chamberlain, 1996, 2008; Gaschnig et al., 2013). Consequently, the East Kootenay tectono-magmatic event has been interpreted as a period of either orogenic collapse (McMechan and Price, 1982; McFarlane, 2015) or volcanism and renewed subsidence near the end of deposition of the Belt Supergroup (Doughty and Chamberlain, 2010; McFarlane and Pattison, 2000; Pattison and Seitz, 2012). Some 1000 km further north, Verbaas et al., (2018) correlate the ca. 1380 Ma Hart River sills with the Salmon River sills, and suggests that the bimodal magmatic provinces may record rifting in western Laurentia. Ernst et al., (2008) considers the Hart River and Salmon River sills to comprise a single large igneous province (LIP) that may record rifting of Nuna. Recently, an  $\geq 2.5$  km thick package of fine-grained and carbonate rocks (the Deer Trail Group) has been shown to be younger than ca. 1300 and may record deposition during the final stages of this East Kootenay tectono-



magmatic event (Box et al., 2020; Brennan et al., 2021b). Remnants of Deer Trail Group strata are likely present elsewhere along the western edge of the Belt Basin (Pearson and Link, 2021), suggesting that a substantial thickness of sedimentary rocks may have been removed from atop the Belt Basin prior to Rodinian rifting after ca. 760 Ma (Brennan et al., 2021a).



**Fig. 4.3:** Geologic map of the U.S. and southern Canadian North American Cordillera, including the major metamorphic core complexes (in blue) and the extent of ca. 1480–1320 Ma magmatism (in red hashed) and the extent of Cordillera associated plutonism (in pink). Inset map shows the Belt basin region in greater detail, including the Priest, River, Clearwater, and Bitterroot core complexes. The extent of the Clearwater regional geologic map, in Fig. 4, is indicated in the inset map.

The younger event is often attributed to ca. 1115–1060 Ma, Grenville-age, tectonothermal activity. This Grenville-age metamorphism has been suggested to simply reflect static magmatic heating at depth, perhaps suggesting a relation to coeval, deep (possibly plume related) magmatism within the Midcontinent Rift, and Southwest Laurentian Large Igneous Province (e.g. Doughty and Chamberlain, 2008; Slack et al., 2020). However, no magmatism of this age has been directly identified within the Belt Basin, while it is relatively widespread in these other coeval provinces. Others have

instead suggested that the ca. 1100 Ma ages instead require tectonic thickening (Zirakparvar et al., 2010; Neishem et al., 2012), similar to the peak of Grenville orogenic burial in southern Laurentia at ca. 1140–1100 Ma (Mulder et al., 2017). Recently, the only suspected ca. 1100 Ma strata in this region, the Buffalo Hump Formation (e.g. Ross et al., 1992), was shown to contain ca. 760 Ma detrital zircon grains. These ages indicate that the Buffalo Hump Formation was actually deposited during Rodinia breakup, not assembly (Brennan et al., 2021a). Consequently, no magmatism or stratigraphy associated with this Grenville-age event within the Belt Basin is identified.

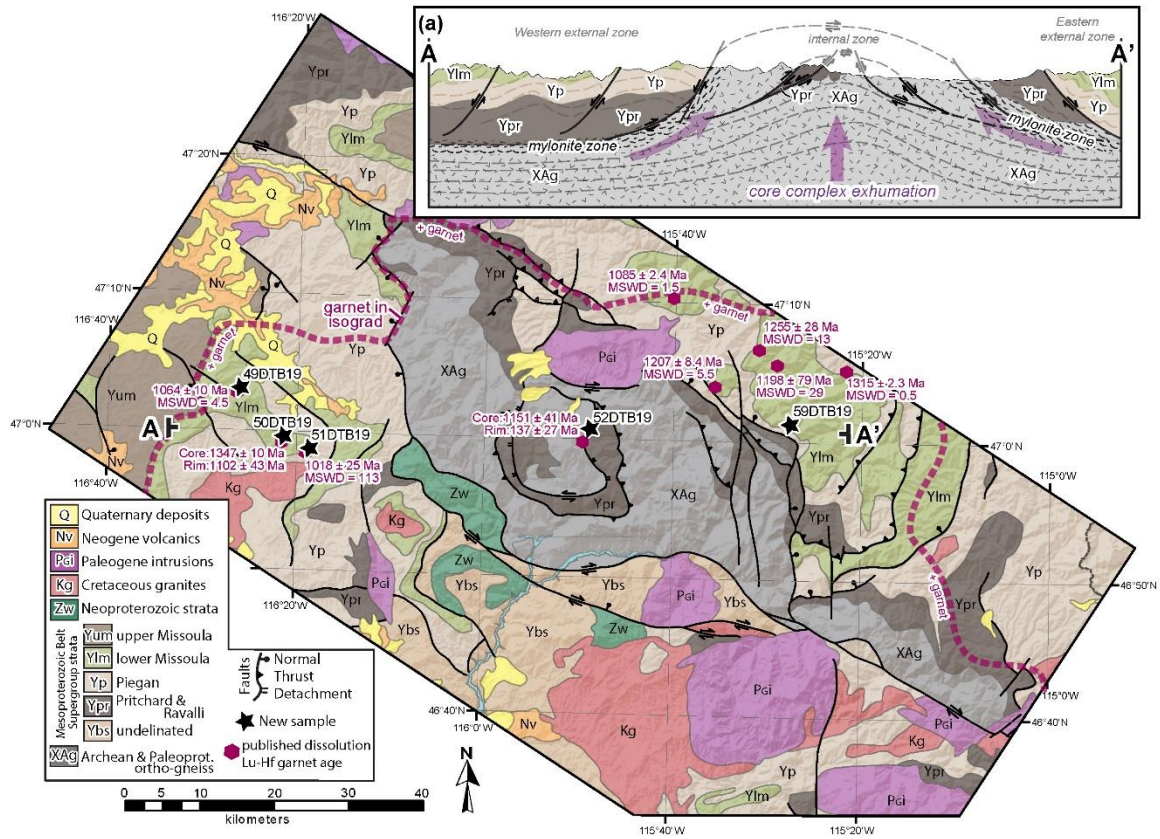
Most Belt Supergroup strata east of the Priest River, Clearwater and Bitterroot complexes record greenschist or lower-grade metamorphism and thus have not been examined by metamorphic studies. However, the enigmatic spread of ca. 1400–1000 Ma ages variably overprinted by younger ca. 130–50 Ma metamorphism recorded within the amphibolite-grade metamorphic assemblages of the within the Priest River, Clearwater and Bitterroot complexes have been investigated. In the Priest River and Bitterroot complexes, metamorphism has mostly been interpreted to relate to late Mesozoic and younger Cordilleran events, consistent with monazite U–Pb ages (Stevens et al., 2015; Aleinikoff et al., 2012). However, structures related to deposition of the Belt Supergroup and ca. 1380–1060 Ma U–Pb cassiterite (SnO<sub>2</sub> ore; Slack, 2020), monazite (McFarlane, 2015) and xenotime (Aleinikoff et al., 2012) ages are known to occur.

Within the Priest River complex, Mesoproterozoic lower Belt Supergroup metapelitic rocks record peak metamorphic conditions of ~785–790 °C and ~9.6–10.3 kbar (constrained by phase equilibria modelling) between ca. 74 and 54 Ma as recorded by U–Pb monazite ages (Stevens et al., 2015). However, approximately 100 km north of the Priest River complex, in a similar structural position within the Cordilleran orogenic belt, lower Belt Supergroup strata in the Matthew Creek metamorphic zone (Fig. 3) are intruded by  $1365 \pm 5$  and  $1335 \pm 5$  Ma peraluminous granites and have an older, pre-Cordilleran metamorphic record. Lower Belt Supergroup metapelites in the Matthew Creek metamorphic zone contain layer-parallel foliation fabrics and peak metamorphic conditions of ~580–650 °C and ~3.5 kbar (also constrained by phase equilibria modelling) that occurred from ca. 1365–1335 Ma, as recorded by U–Pb monazite ages (McFarlane and Pattison, 2000; McFarlane, 2015). Some 10 km to the northeast, ca. 1468 Ma cassiterite hosted within metapelitic rocks of the lower Belt

Supergroup records thermal events (of unknown P/T conditions) at ca. 1380 and 1075 Ma (Slack, 2020).

Along the southern region of the Belt Basin, the Bitterroot complex records peak metamorphic conditions of ~700–750 °C and ~7–8 kbar (based on early garnet and amphibole composition thermobarometry methods; House et al., 1997) between ca. 64 and 56 Ma (Aleinikoff et al., 2012; Howlett et al., 2021). Monazite from Mesoproterozoic rocks in the Bitterroot complex generally record Cretaceous (ca. 110–92 Ma) U–Pb ages that are similar in age to the intrusion of nearby granites from the Idaho Batholith (Aleinikoff et al., 2012; Bookstrom et al., 2016). However, rocks in the Bitterroot complex also record evidence for pre-Cordilleran metamorphism, particularly within the Blackbird mineralized district that contains ca. 1370 Ma bimodal magmatism, ca. 1370–1320 Ma U–Pb xenotime ages, and ca. 1060 Ma hydrothermal xenotime ages (Aleinikoff et al., 2012). Mesoproterozoic metamorphism in this region is estimated to require pressures of 4–6 kbar (based on garnet thermobarometry methods) equating to ~14–20 km of burial (Doughty and Chamberlain, 2010; Bookstrom et al., 2016).

The Clearwater complex (Fig. 4.4) also records peak metamorphic conditions of ~700–750 °C and pressures as high as ~9 kbar (based on garnet thermobarometry methods) between ca. 82 and 64 Ma constrained by U–Pb ages of metamorphic zircon rims (Doughty et al., 2007). Doughty et al. (2007) only looked at the internal zone of the complex, which was likely buried the deepest in the Cordilleran orogeny, and suggests that the Clearwater complex experienced similar Cordilleran metamorphism as the previously described complexes to the north and south. However, early workers (Lang and Rice, 1985) recognized that metamorphism within the internal zone overprinted an early regional metamorphic event that is better preserved in the external (marginal) zones of the complex (Fig. 4.4; Doughty et al., 2007; Baldwin et al., 2016). Subsequent Lu–Hf garnet geochronology on metapelitic rocks primarily within these external zones revealed an enigmatic spread of ca. 1400–1000 Ma ages (Zirakparvar et al., 2010; Neishem et al., 2012) with no record of late Mesozoic to early Paleogene metamorphism (Doughty et al., 2007). These results suggest that the external zone of the Clearwater complex may contain a better preserved record of these Mesoproterozoic metamorphic events than amphibolite-grade regions to the north (Priest River complex and Matthew Creek metamorphic zone) or south (Bitterroot complex and Blackbird district).



**Fig 4.4:** Geologic map of the Clearwater complex (modified from Lewis et al., 2012), including the published ca. 1350–1020 Ma garnet Lu–Hf ages in this region (Zirakparvar et al., 2010; Neishem et al., 2012), and the location of our samples along the western external region of the complex (49, 50, 51DTB19), center (52DTB19), and eastern external region (59DTB19). Schematic core complex cross section cartoon (A–A’) after Doughty et al., (2007).

### 4.3. Methods

#### 4.3.1 Mineralogy

From selected rock billets, Wagner Petrographic cut and polished (30  $\mu\text{m}$  thick) thin sections for petrographic investigation. The thin sections were investigated using standard transmitted light microscopy and TESCAN integrated mineral analyzer (TIMA) at the John de Laeter Centre at Curtin University, Perth Australia. TIMA mapping of polished thin sections provided detailed X-ray and back-scattered electron (BSE) maps for mineralogical characterization, including quantification of mineral modes and the identification of accessory phases for subsequent dating. TIMA analyses used a 15mm working distance, a 2500eV beam intensity and a spot size of 50- $\mu\text{m}$ . Full thin section photomicrographs were taken with a Zeiss Axio Imager 2 imaging system. A JEOL 8530F electron probe microanalyser (EPMA) at the Centre for Microscopy, Characterization and Analysis, University of Western Australia was

used to measure semi-quantitative garnet mineral compositions. EPMA parameters included a 15kV accelerating voltage, 20 nA beam current, 100-ms dwell time and utilized natural and synthetic standards. Garnet formula were calculated stoichiometrically based on 12 oxygen following Droop (1987).

#### **4.3.2 Phase equilibrium modelling**

Material from the same hand sample as used for the thin sections was analyzed for bulk rock geochemistry by the Bureau Veritas Laboratory, in Perth, Western Australia using X-Ray Fluorescence (XRF). Isochemical  $P$ – $T$  phase diagrams (pseudosections) were generated for the major element oxide compositions using THERMOCALC 3.50 (Holland and Powell, 2011) and the activity–composition models of White et al., (2014a, 2014b) within the MnO–Na<sub>2</sub>O–CaO–K<sub>2</sub>O–FeO–MgO–Al<sub>2</sub>O<sub>3</sub>–SiO<sub>2</sub>–H<sub>2</sub>O–TiO<sub>2</sub>–O (MnNCKFMASHTO) system. Calculations used an Fe<sup>3+</sup>/Fe<sup>2+</sup> ratio of 0.2 (Johnson et al., 2021).

#### **4.3.3 Laser-ablation mass spectroscopy**

Monazite and apatite grains were targeted in-situ within the thin sections for measurement of U–Pb isotopes. During this analysis, material ablated by a Resonetics RESolution SE 193nm laser incorporating a dual volume S155 sample cell was sent to an Agilent 8900 triple-quadrupole inductively coupled mass spectroscopy (LA-QQQ-ICPMS) which measured <sup>202</sup>Hg, <sup>204</sup>Pb (and <sup>204</sup>Hg), <sup>206</sup>Pb, <sup>207</sup>Pb, <sup>208</sup>Pb, <sup>232</sup>Th, and <sup>238</sup>U masses. All data was collected across two analytical sessions at the GeoHistory Facility in the John de Laeter Centre at Curtin University, Perth Australia. In Iolite 3.5, the monazite time-resolved mass spectra were reduced using the U\_Pb\_Geochronology4 data reduction scheme (Paton et al, 2011). Due to the presence of common-Pb bearing standards, apatite time-resolved mass spectra were reduced using VizualAge\_UcomPbine reduction scheme (Chew et al., 2014 and references within). The VizualAge\_UcomPbine reduction scheme applies a common-Pb correction to the primary standard before applying a drift correction to the unknowns; otherwise, no common (or non-radiogenic) Pb-correction was applied. Based on the long term-reproducibility of U-Pb standard analysis at the GeoHistory Facility an additional 2% error was propagated to individual analyses.

Due to the variable common-Pb content of apatite, the uncorrected U–Pb apatite data was plotted on Terra-Wasserburg Concordia diagrams (<sup>207</sup>Pb/<sup>206</sup>Pb vs. <sup>238</sup>U/<sup>206</sup>Pb), and the apatite age was calculated as the lower Concordia intercept age of a free-regression Discordia line, with the common Pb-component represented by the



y-intercept on the  $^{207}\text{Pb}/^{206}\text{Pb}$  axis (Chew et al., 2011; Kirkland et al., 2018). This approach assumes that the U–Pb data are concordant and equivalent. To a lesser degree, monazite can also sometimes include a component of common Pb. Where a common-Pb component was apparent within some monazite analyses, a similar approach was employed which is covered further in the discussion. All data was calculated and plotted using IsoplotR (Vermeesch, 2018).

For monazite U–Pb analysis, monazite grains were targeted with a 10  $\mu\text{m}$  laser spot. Monazite 44069 was utilized as the primary standard and gave a weighted mean age of  $424.1 \pm 4.6$  Ma (mean square weighted deviation or MSWD = 0.02) which is the same as the accepted value ( $424.9 \pm 0.4$  Ma; Aleinikoff et al., 2006). Analysis of secondary reference standards, Manangotry ( $558 \pm 3$  Ma; Horstwood et al., 2003) and Stern ( $511.7 \pm 1.2$  Ma; Palin et al., 2013) gave weighted mean ages ( $562.4 \pm 7.6$ , MSWD = 0.3;  $509.9 \pm 5.42$ , MSWD = 0.38) within uncertainty of accepted values. Only monazite U–Pb analyses that overlap Concordia within  $2\sigma$  uncertainties are defined as concordant (e.g. Spencer et al., 2016).

Apatite U–Pb analyses used a 30  $\mu\text{m}$  laser spot and data were calibrated against the Madagascar apatite standard, which gave a weighted  $^{238}\text{U}/^{206}\text{Pb}$  mean age of  $472.8 \pm 5.1$  Ma (MSWD = 0.06) which is the same as the published value ( $474.25 \pm 0.85$ ; Thomson et al., 2012). Secondary reference standards Forest Center–FC ( $1094 \pm 34$ ; Thomson et al., 2012) and McClure Mountain–MMC ( $523.5 \pm 1.47$  Ma; Schoene and Bowring, 2006), when treated as unknowns and allowed free regression on a discordia line in Tera-Wasserburg space yielded lower concordia intercept ages ( $1098.9 \pm 34.2$ , MSWD = 0.39 and  $522.8 \pm 17.3$ , MSWD = 0.92, respectively), also consistent with published values.

## **4.4. Results**

### **4.4.1 Field and petrographic observations**

The metapelitic samples (49, 50, 51, and 59DTB19) were all collected from the Wallace Formation of the Belt Supergroup (Fig. 4.4; Lewis et al., 1999; 2001; 2005). To the north and east, where the Wallace Formation is only weakly metamorphosed, it consists mostly of fine sandstone and siltstone, commonly in graded couplets, with dolomite and carbonate intervals, and is overlain by at least 8 km of upper Belt Supergroup strata (Lonn et al., 2020). Approximately 250 km northeast of the study area within the Clearwater complex, a volcanic tuff from a

similar stratigraphic position yielded a  $1454 \pm 9$  Ma age, and a stratigraphically-higher rhyolitic lava an age of  $1443 \pm 7$  Ma age (Evans et al., 2000), consistent with a ca. 1450 Ma depositional age for the protoliths of the metapelitic rocks discussed below.

*Western extent of Clearwater Complex—Samples 49, 50, and 51DTB19*

Sample 49DTB19 is a garnet–muscovite–staurolite–biotite–quartzofeldspathic schist with relict (4 to 20 mm thick) graded bedding (Fig. 4.5). This location was previously sampled (sample 05JV07) by Zirakparvar et al., (2010). They report a Lu–Hf garnet age of  $1064 \pm 10$  (MSWD = 4.5) from this locality. In our collected sample (49DTB19), a foliation is mostly defined by alignment of feldspar and micas and is mostly subparallel to bedding. Staurolite porphyroblasts are up to 4 cm long and may be twinned. Garnet forms small (mostly <1 mm) porphyroblasts containing apparently randomly-oriented inclusions (primarily quartz) that are preferentially concentrated in the cores of grains. There is no observed deflection in (i.e. wrapping of) the foliation around garnet or staurolite porphyroblasts (Fig. 4.5A, inset i, inset ii; Fig. 4.6A). Monazite and apatite are present within the matrix but neither occur as inclusions within garnet, and neither garnet, monazite or apatite are found as inclusions within staurolite.

Sample 50DTB19 is a staurolite–muscovite–garnet–biotite–sillimanite–quartzofeldspathic schist from the same location of sample 08HL-04 of Neishem et al., (2012), who report similar lithologies and Lu–Hf garnet dissolution ages of  $1347 \pm 10$  Ma (MSWD = 4.6; garnet cores) and  $1102 \pm 43$  Ma (MSWD = 42; garnet rim). Sillimanite comprises both large prismatic porphyroblasts (up to 30 mm long; Fig. 4.5B, inset i), and as fibrolite (Fig. 4.5B, inset ii). The foliation is defined primarily by matrix micas, and sillimanite porphyroblasts. In thin section, thin veinlets filled primarily with K-feldspar crosscut the prismatic sillimanite porphyroblasts (Fig. 4.6B). Small, euhedral to subhedral staurolite crystals are found along the margins of sillimanite (Fig. 4.6B). Garnet grains within 50DTB19 are mostly euhedral, up to 3 mm in diameter, and are clearly wrapped by the foliation, locally exhibiting evidence for dissolution and quartz strain shadows. Garnet grains generally show an inclusion-rich (mostly quartz) core. Monazite grains occur within the matrix and as inclusions within garnet.

Sample 51DTB19 is a quartzofeldspathic muscovite–garnet–biotite–sillimanite schist. Sillimanite is present as small (up to 2 mm long) clusters of

fibrolite that are preferentially aligned parallel a subtle foliation and are crosscut by K-feldspar (Fig. 4.6C). The mica- and quartz-defined foliation within this sample is less strongly developed and wraps large ( $\geq 5$  mm in diameter) garnet porphyroblasts (Fig. 4.5Ci). This sample is from the same location as sample 05JV03 of Zirakparvar et al., (2010), who report similar lithologies and a Lu–Hf garnet dissolution age of  $1018 \pm 25$  Ma (MSWD = 113). Based on the high MSWD of the Lu–Hf isochron including all five garnet fractions analysed, Zirakparvar et al., (2010) propose discrete Lu–Hf isochron model ages of  $1024 \pm 3$  Ma (MSWD = 1.5) and  $1081 \pm 20$  Ma (MSWD = 45) based on inclusion-richer and inclusion-poorer garnet fractions.

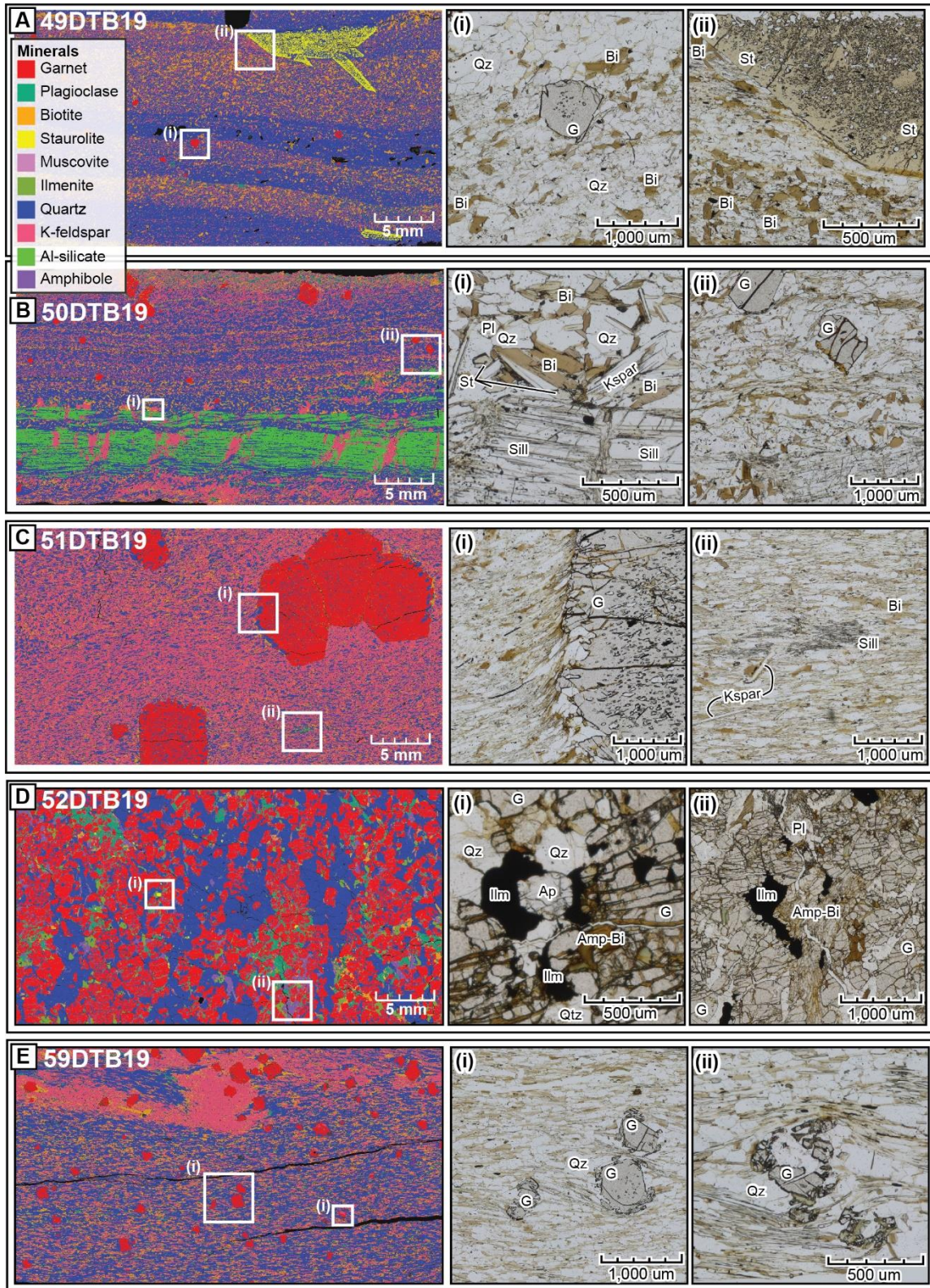
#### *Core of Clearwater Complex–Sample 52DTB19*

Sample 52DTB19 is a strongly-lineated mylonitic ilmenite- and apatite-bearing garnet amphibolite from the central portion of the Clearwater complex. In thin-section, interstitial quartz is abundant, suggesting alternation by hydrothermal fluids (Fig. 4.5D), and consequently this sample was not utilized for thermodynamic modelling, but was previously analyzed for zircon U/Pb and O- isotopes (Brennan et al., 2022), which we supplement with new apatite U/Pb results in this study.

#### *Eastern extent of Clearwater Complex–Sample 59DTB19*

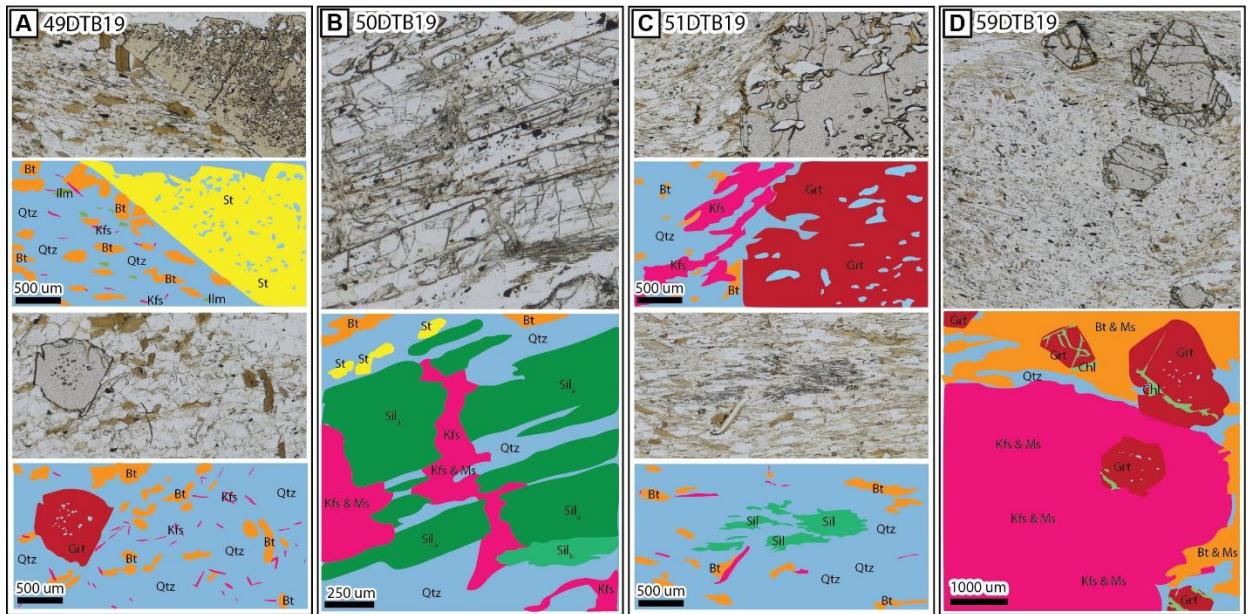
Sample 59DTB19 is a quartzofeldspathic monazite- and apatite-bearing muscovite–garnet–biotite schist, collected along the eastern extent of the complex. Although no garnet ages are reported from this specific outcrop, 10 to 22 km to the north, Neishem et al., (2012) reports Lu–Hf garnet ages that range from ca. 1315 to 1085 Ma (see Fig. 4.4). In this sample, the foliation is mostly defined by quartz, feldspar and micas. Garnet grains have inclusion-rich (mostly quartz) cores and exhibit dissolution textures. Garnet grains locally show core/rim geometries and exhibit development of strain shadows that are filled with biotite, muscovite, and quartz (Fig. 4.5E inset i and ii; Fig. 4.6DE). Monazite and apatite grains are present within the matrix. A couple of monazite grains are also present as inclusions within garnet.





**Fig. 4.5:** Representative thin section mineralogy (produced by TESCAN integrated mineral analyzer; TIMA) and important mineralogical relationship highlighted inset boxes (i) and (ii). Mineral abbreviations from Whitney and Evans (2010).





**Fig. 4.6:** Additional mineralogy of the metapelitic samples shown in plane polarized light microphotographs, and interpretative sketches. (A) Sample 49DTB19 garnet-muscovite-staurolite-biotite-quartzofeldspathic schist. (B) Sample 50DTB19 staurolite-muscovite-garnet-biotite-sillimanite-quartzofeldspathic schist, highlighting the two generations of sillimanite and their association, but non-cross cutting or inclusion, relationship with staurolite. (C) Sample 51DTB19 muscovite-garnet-biotite-sillimanite-quartzofeldspathic schist, showing the fibrolite sillimanite present, and large garnets. (D) Sample foliated monazite and apatite bearing muscovite-garnet-biotite-quartzofeldspathic schist, showing a large feldspar and muscovite-rich region that is perhaps a pseudomorph of an earlier higher-grade assemblage. Mineral abbreviations from Whitney and Evans (2010).

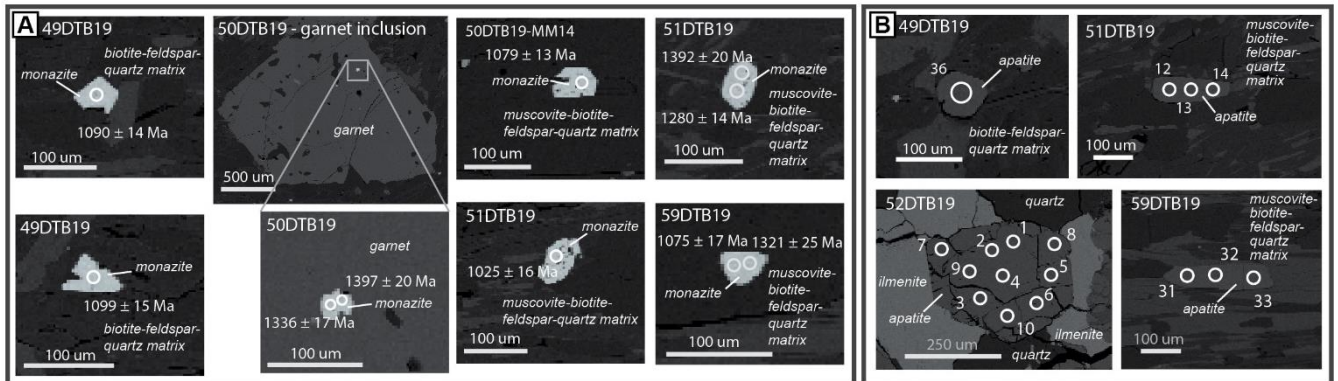
#### 4.4.2 U–Pb Geochronology

##### *Western extent of Clearwater Complex (samples 49, 50, and 51DTB19)*

Monazite from the analyzed samples (Fig. 4.7A) yielded concordant U–Pb dates between ca. 1400 and 1060 Ma (Fig. 4.8). Apatite U–Pb results indicated younger dates, and generally fall into either a Cryogenian (ca. 676 Ma) or late Paleogene-early Eocene (ca. 58–54 Ma) population.

Monazite within sample 49DTB19 only occurs within the matrix. A free Discordia regression line fit through 15 analyses yields a  $1074 \pm 14$  (MSWD = 0.6) age, with a  $^{207}\text{Pb}/^{206}\text{Pb}$  common Pb upper intercept of  $\sim 0.9$ . Ten of the 15 analyses are concordant, and together yield a concordia age of  $1087 \pm 10$  (MSWD = 1.9). A free Discordia regression line fit through 38 analyses of matrix apatite in sample 49DTB19 (Fig. 4.7B) yields a lower intercept age of  $676 \pm 13$  Ma (MSWD = 2.8), and an upper  $^{207}\text{Pb}/^{206}\text{Pb}$  common Pb intercept of  $\sim 0.9$  (Fig. 4.9). Zirakparvar et al.,

(2010) report a  $1064 \pm 10$  Ma (MSWD = 4.5) garnet Lu–Hf age for the same outcrop.



**Fig. 4.7** (A) Representative back-scattered electron (BSE) images of U–Pb analyzed monazite with concordant ages labelled. Note that samples 50DTB19 (shown) and 51DTB19 contained monazite grains as inclusions within garnet that yielded older ca. 1365–1320 Ma ages. Of the limited monazite grains present in 59DTB19, a single grain able to fit two analyses spots yielded both ca. 1321 and 1075 Ma ages. (B) Representative back-scattered electron (BSE) images of U–Pb apatite grains with individual analyses spots indicated that correspond to the “Spot ID” within the supplementary tables. Note individual Pb-corrected apatite grain ages were not calculated and consequently not labelled here.

In sample 50DTB19, monazite occurs both in the matrix and as inclusions within garnet. Five of six monazite grains analyzed as inclusions within garnet show concordant dates (Fig. 4.7A; 4.8) that yield a concordia age of  $1365 \pm 17$  Ma (MSWD = 0.1). Twenty-one analyses of matrix monazite yield 11 concordant dates ranging from  $1358 \pm 15$  to  $1060 \pm 14$  Ma (all individual spot ages reported as single spot concordia ages; Ludwig, 1998), the two youngest of which are  $1079 \pm 13$  and  $1060 \pm 14$  Ma. The two youngest concordant analyses, and the five youngest discordant grains with a common Pb anchor (of 0.9, e.g. Albarede and Martine, 1984) define a Discordia with an intercept age of  $1056 \pm 19$  Ma (Fig. 4.8). Sample 50DTB19 does not contain any apatite. Neishem et al., (2012) microdrilled (rim and core zones) garnets from this same sampled outcrop and found Lu–Hf ages of  $1347 \pm 10$  Ma (MWSD = 4.6; isochron from garnet cores) and  $1102 \pm 43$  Ma (MSWD = 42; isochron from garnet rims).

Sample 51DTB19 contains monazite within the matrix and as inclusions within garnet (Fig. 4.7A). Seven of nine monazite grains were analyzed as inclusions within garnet yielded concordant dates that range from  $1392 \pm 20$  to  $1269 \pm 19$  Ma, and collectively give a concordia age of  $1318 \pm 14$  Ma (MSWD = 0.3). A single inclusion monazite grain sufficiently large for two analytical spots yields different concordant ages ( $1392 \pm 20$  and  $1280 \pm 14$  Ma; Fig. 4.7). Twenty-six of 30 concordant matrix monazite grains yield dates that range from  $1379 \pm 21$  to  $1007 \pm$

17 Ma. The three youngest monazite analyses yield a weighted mean age of  $1021 \pm 20$  Ma (MSWD = 3.7; Fig. 4.8). For this location, Zirakparvar et al., (2010) reported a  $1018 \pm 25$  Ma (MSWD = 113) Lu–Hf isochron age based on five garnet fractions (i.e. five fractions of a single garnet). Given the high MSWD, Zirakparvar et al., (2010) speculate their data may represent two stages of garnet growth at  $1081 \pm 20$  (MSWD = 45, three garnet fractions) and  $1024 \pm 3$  (MSWD = 1.4, two garnet fractions). Analyses of 32 matrix apatite grains (Fig. 4.7B) define a free Discordia regression with a lower intercept age of  $56 \pm 7$  Ma (MSWD = 22), and an upper  $^{207}\text{Pb}/^{206}\text{Pb}$  common Pb intercept of  $\sim 0.48$  (Fig. 4.9). The high MSWD of the Discordia regression indicates complexity in the dataset, such as potentially multiple age-populations, which will be explored further in the discussion.

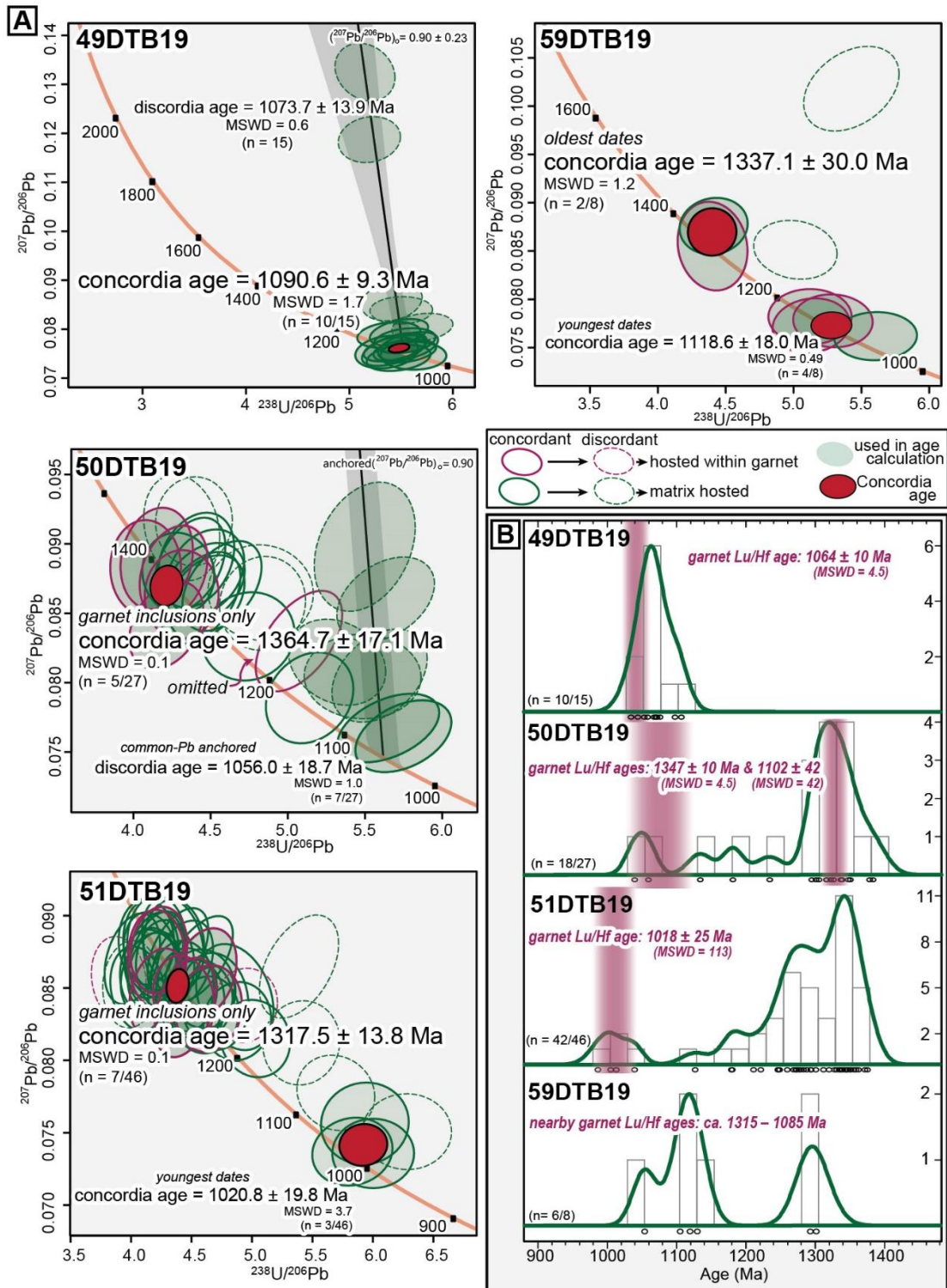
*Core of Clearwater Complex (sample 52DTB19)*

Sample 52DTB19 lacks monazite but contains abundant apatite (Fig. 4.7B). A free Discordia regression line fit through 91 apatite analyses yields a lower intercept age of  $54 \pm 2.7$  Ma (MSWD = 1.3), and an upper  $^{207}\text{Pb}/^{206}\text{Pb}$  common Pb intercept of  $\sim 0.50$  (Fig. 4.9). Brennan et al., (2022) report zircon U–Pb and O-isotope data from this sample that indicate metamorphic/hydrothermal zircon growth at ca. 1396–1350 Ma.

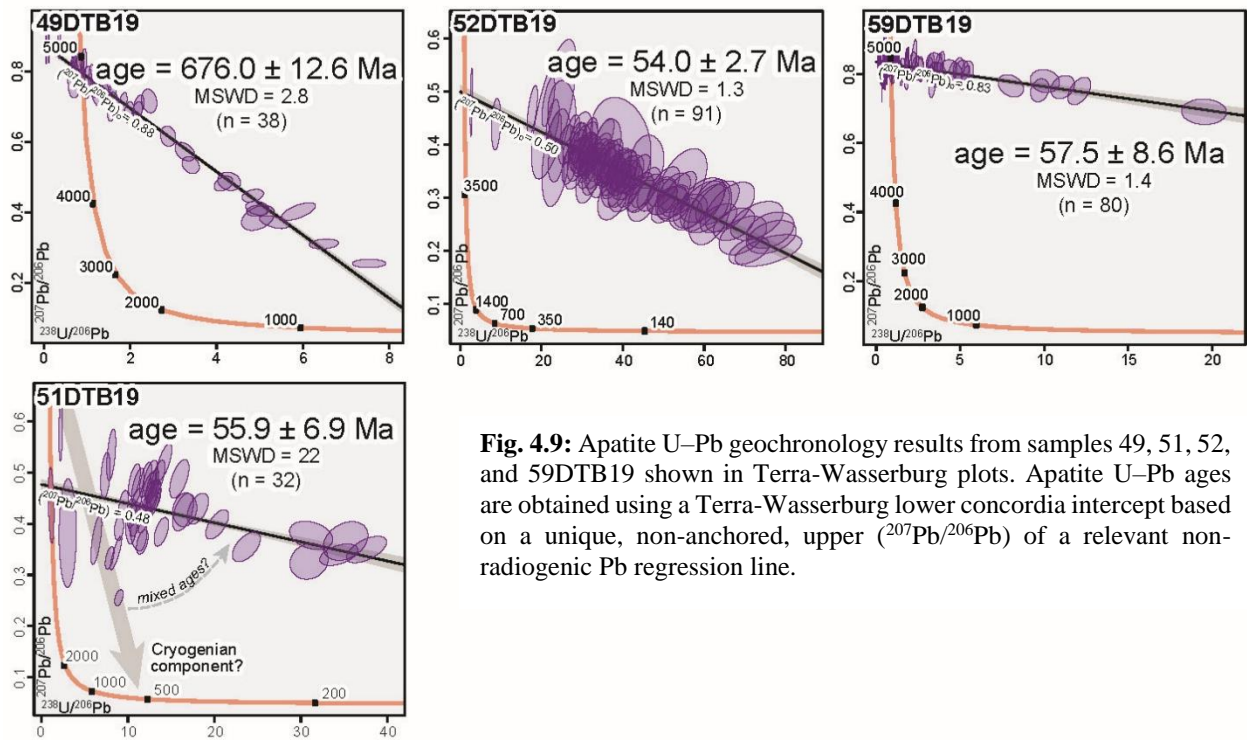
*Eastern extent of Clearwater Complex–Sample 59DTB19*

Sample 59DTB19 contains rare, both garnet and matrix hosted, monazite (Fig. 4.7A). Six of eight concordant analyses give dates that fit into two age-populations, one at around 1337 Ma and another at around 1119 Ma (Fig. 4.8A). A single grain sufficiently large for two spot analyses yields concordant ages of ca. 1321 and 1075 Ma. The older two concordant analyses give dates of  $1321 \pm 25$  and  $1346 \pm 19$  Ma, that together define a concordia age of  $1337 \pm 30$  Ma (MSWD = 1.2). The younger four concordant analyses dates range from  $1150 \pm 20$  to  $1075 \pm 17$  Ma, and give a concordia age of  $1119 \pm 18$  Ma (MSWD = 0.5; Fig. 4.8). A discordia through 80 analyses of matrix apatite (Fig. 4.7B) define a lower intercept age of  $58 \pm 9$  Ma (MSWD = 1.4), and an upper  $^{207}\text{Pb}/^{206}\text{Pb}$  common Pb intercept of  $\sim 0.8$  (Fig. 4.9).





**Fig. 4.8:** Monazite U–Pb geochronology results from samples 49, 50, 51, and 59DTB19. Results are shown in (A) Terra-Wasserburg plots, please note the legend for these plots beneath sample 59DTB19 and (B) Kernel Density Plots (plotted with a 25 Myr band and binwidth) which include the corresponding published garnet Lu–Hf dissolution ages (Zirakparvar et al., 2010; Neishem et al., 2012). Only analyses that overlap Concordia within  $2\sigma$  uncertainties are plotted. Monazite grains hosted along cracks within garnet are plotted here as matrix hosted but are indicated in Appendix D4.4.



**Fig. 4.9:** Apatite U–Pb geochronology results from samples 49, 51, 52, and 59DTB19 shown in Terra-Wasserburg plots. Apatite U–Pb ages are obtained using a Terra-Wasserburg lower concordia intercept based on a unique, non-anchored, upper ( $^{207}\text{Pb}/^{206}\text{Pb}$ ) of a relevant non-radiogenic Pb regression line.

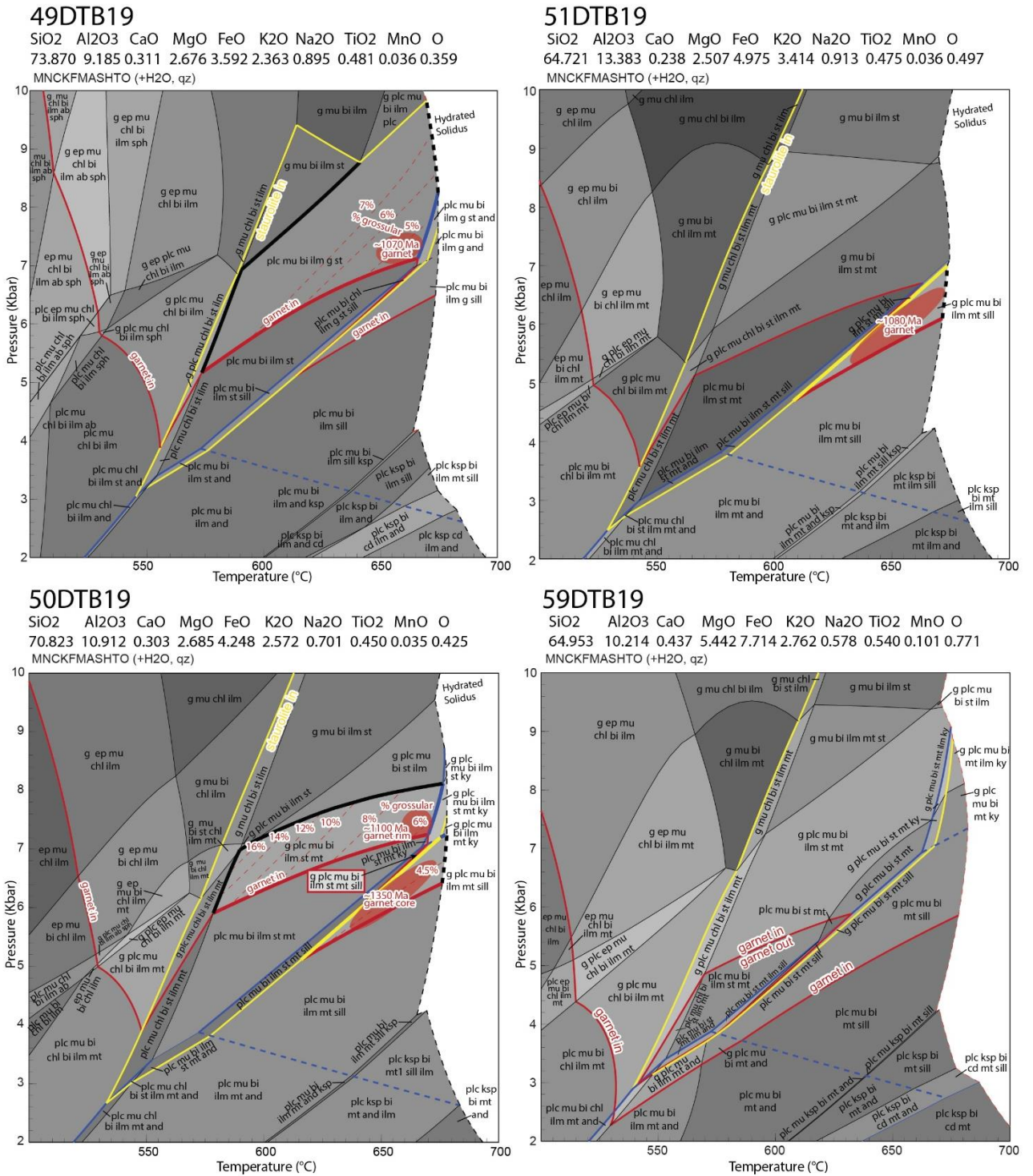
#### 4.4.3 Mineral Compositions and Phase Equilibrium Modelling

Pressure–temperature pseudosections for the four modelled samples are shown in Fig. 4.10, which the inferred peak assemblage fields are highlighted. For clarity, we show the stability of garnet (at P/T above the thick red line) and staurolite (at P/T above the thick yellow line).

##### *Western extent of Clearwater Complex*

The three samples collected along the western region of the Clearwater have similar major oxide bulk compositions and, consequently, similar calculated phase equilibria. None of the samples preserves evidence for partial melting, restricting peak metamorphic conditions to temperatures below the H<sub>2</sub>O-saturated solidus (< or << 680 °C; Fig. 4.10).

Sample 49DTB19 contains an interpreted peak assemblage of garnet–muscovite–staurolite–biotite–plagioclase–ilmenite (+quartz, and H<sub>2</sub>O), consistent with conditions of ~570–670 °C and 5.5–9.0 kbar. Garnet within this sample shows flat major element profiles with around 3 mol.% spessartine, 5 mol.% grossular (Ca<sub>3</sub>Al<sub>2</sub>), 11 mol.% pyrope, and 80% mol. almandine, and an Fe# = 89% (where Fe# = Fe<sup>2+</sup>/[Fe<sup>2+</sup> + Mg]; Fig. 4.11). As Ca is usually the slowest diffusing major cation (Carlson, 2006), we use the calculated 5 mol. grossular isopleth to more tightly constrain peak P–T conditions to 6.5–7.5 kbar, and 640–665 °C (Fig. 4.10).

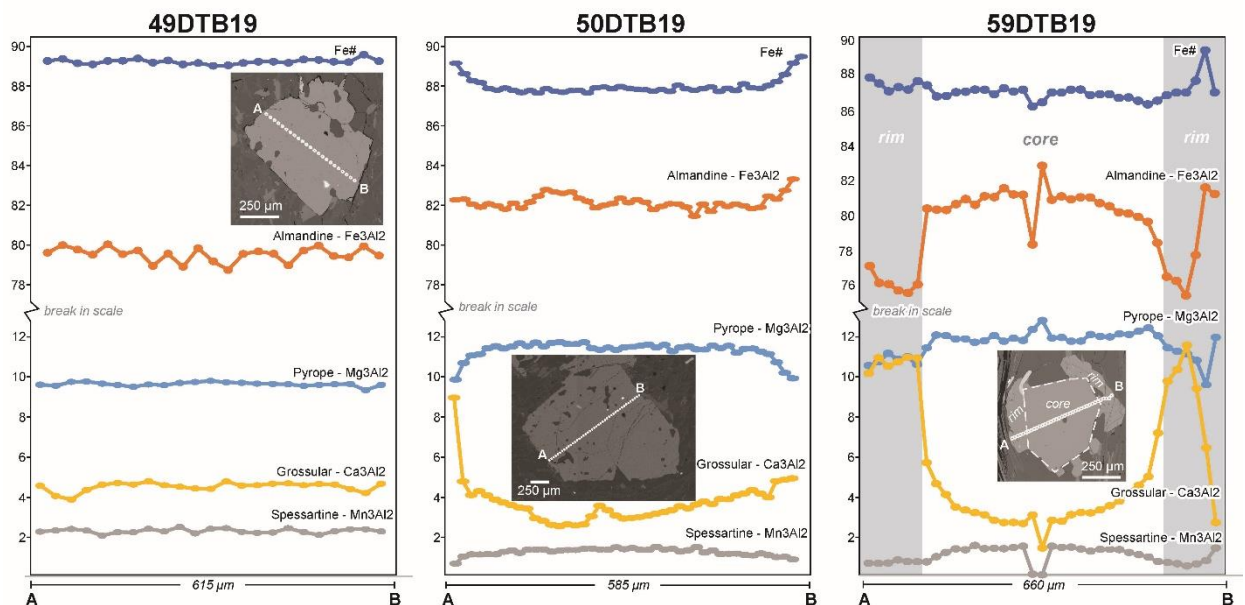


**Fig. 4.10:** Pressure–temperature pseudosections calculated for 49, 50, 51, and 59DTB19. The garnet-in isopleth is indicated in red, staurolite-in isopleth is indicated in yellow, and aluminosilicate triple point fields indicated in blue. Fields containing the observed mineral assemblages are highlighted with thick lines and the interpreted peak paragenesis is indicated by the red field. Please note for sample 50DTB19, the observed mineral assemblage falls within the narrow field labelled by the red box, while the stability field indicated by the garnet chemistry are shown in the red ovals. Mineral abbreviations from Whitney and Evans (2010). H<sub>2</sub>O and quartz (qz) were considered in excess for all calculations.



Sample 50DTB19 contains the interpreted peak assemblage garnet–muscovite–staurolite–biotite–plagioclase–ilmenite–sillimanite (+H<sub>2</sub>O and quartz; Fig. 4.5), which is predicted to be stable within a thin field from ~620 °C and 5 kbar, to ~680 °C and ~6.8 kbar (Fig. 4.10). The core of the garnets, dated at 1347 ± 10 Ma (Zirakparvar et al., 2010) show ~3–4.5% grossular (Fig. 4.11), which is broadly consistent with a modelled lower-pressure assemblage of garnet–muscovite–biotite–plagioclase–ilmenite–magnetite–sillimanite (+H<sub>2</sub>O and quartz) stable from ~620 °C and ~5.2 Kbar to ~680 °C and 7 Kbar (Fig. 4.10). The garnet rims, dated at 1102 ± 42 Ma (Zirakparvar et al., 2010), show a slight increase in percent grossular to perhaps as high as 9% grossular (Fig. 4.11). This compositional change in garnet is consistent with slightly higher-pressure garnet growth, and the assemblage of garnet–muscovite–staurolite–biotite–plagioclase–ilmenite–magnetite (+H<sub>2</sub>O and quartz), which is stable from ~6.8–7.8 Kbar and ~650–670 °C, indicating a very similar P/T condition to sample 49DTB19.

Sample 51DTB19 contains the interpreted peak assemblage garnet–muscovite–biotite–plagioclase–ilmenite–magnetite–sillimanite (+quartz, and H<sub>2</sub>O). The absence of staurolite is consistent with slightly lower pressure conditions than recorded within samples 49DTB19 and 50DTB19. Sillimanite and garnet are only predicted to be stable together in a small field between ~620 °C and 5 kbar to ~680 °C and ~6.8 kbar (Fig. 4.10).



**Fig. 4.11:** Garnet compositional profiles of major elements for samples 49, 50, and 59DTB19 including back-scattered electron (BSE) images of analyzed traverses.



### *Eastern extent of Clearwater Complex*

The calculated stability field for garnet in sample 59DTB19, from the eastern extent of the Clearwater complex, is significantly expanded compared to the other samples (Fig. 4.10), reflecting a threefold increase in the Mn content (e.g. White et al., 2014b). A small triangular garnet-absent field is also present related to the stability of staurolite and biotite at the expense of garnet and chlorite. Distinct rim and core domains (Fig. 4.11) suggest at least two stages of garnet growth. The preserved assemblage in this sample lacks staurolite, any aluminosilicate minerals, and contains significant potassium feldspar and muscovite agglomerates, that may have replaced an earlier higher-grade assemblage. Consequently, we interpret significant overprinting by lower (sub-garnet) grade metamorphism in this sample and do not assign a peak assemblage field.

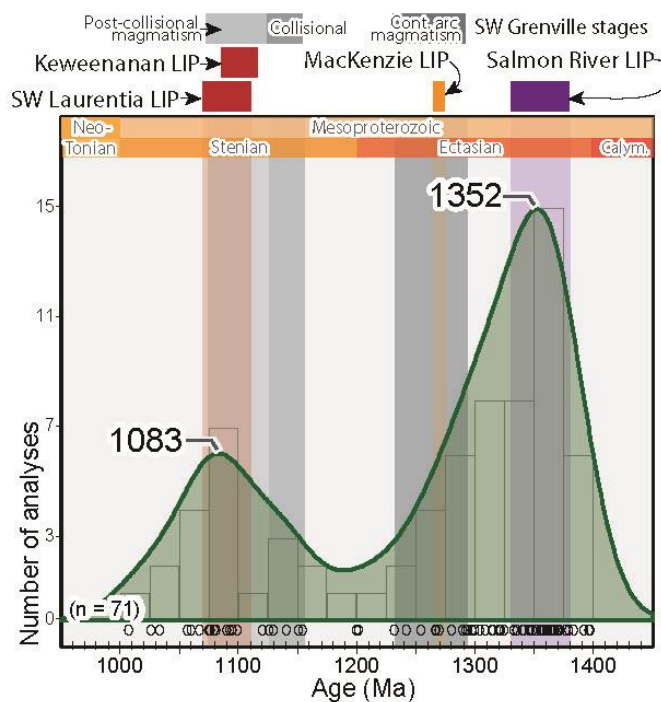
## **4.5. Discussion**

### **4.5.1 Chronological constraints on Metamorphism within the Clearwater region**

Within the external regions (three samples from the western region, one from the eastern) of the Clearwater complex, the ages of 71 concordant monazite U–Pb analyses range from 1397 to 1007 Ma (Fig. 4.12), similar to the range of Lu–Hf garnet ages from the same rocks (1379 to 1018 Ma; Zirakparvar et al., 2010; Neishem et al., 2012). Lu–Hf dates from garnet are usually interpreted to represent its growth, which in metapelitic rocks generally requires pressures greater than 5 kbar, but can occur at temperatures as low as ~450 °C (Yakymchuk et al., 2017). Net diffusive gain or loss of Lu or Hf from garnet is thought to be negligible at  $T < 800\text{--}900$  °C (Smit et al., 2013; Bloch et al., 2020). Monazite in metapelitic rocks can also grow at temperatures as low as ~450 °C, and become closed to net diffusive gain or loss of U or Pb isotopes at  $T < 725$  °C (Parrish, 1990) or higher (e.g. Cherniak, 2010). The U–Pb in apatite system is thought to have a much lower closure temperature of 450–550 °C, making it a valuable tool for constraining middle- $T$  (mid-crustal) processes (Kirkland et al., 2018; Jepson et al., 2021).

The concordant monazite U–Pb analyses from the four studied samples indicate two age-peaks at 1352 Ma and 1083 Ma (Fig. 4.12), consistent with growth during the ca. 1380–1300 Ma East Kootenay and ca. 1115–1065 Ma Grenvillian tectono-thermal events (Zirakparvar et al., 2010; Neishem et al., 2014). On an

individual sample basis, the northwestern most sample (49DTB19) records the strongest Grenville-age signature with a monazite U–Pb Concordia age of  $1091 \pm 9$  Ma (MSWD = 1.7), slightly older than the  $1064 \pm 10$  Ma (MSWD = 4.5) Lu–Hf garnet age (Zirakparvar et al., 2010; Fig. 4.8B). Interestingly, this is the only sample to contain a robust pre-Cordilleran U–Pb apatite age of  $676 \pm 13$  Ma (MSWD = 2.8), contemporaneous with significant rift-related volcanism within the ca. 685–650 Ma Edwardsburg volcanics and Big Creek plutonic suite some 150 km to the south (Lund et al., 2003; 2010). Importantly, the absence of Phanerozoic ages indicates that this sample was largely unaffected by Cordilleran metamorphism (implying  $T < 450$  °C).



**Fig. 4.12:** All concordant monazite U–Pb ages from the Clearwater complex, plotted with a 25 Myr band and binwidth. The ages of corresponding southwestern Grenville stages (Mulder et al., 2017), and notable Laurentian Large Igneous Provinces (LIPs; Ernst et al., 2008 and references within) are indicated.

Sample 50DTB19 records a wider spread of monazite U–Pb dates, with five monazite grains armored as inclusions in garnet yielding a  $1365 \pm 17$  Ma (MSWD = 0.1) concordia age, with the youngest seven grains defining a  $1056 \pm 19$  Ma (MSWD = 1.0) discordia intercept age. These ages are in good agreement with the Lu–Hf garnet ages in rocks from the same outcrop, which records two stages of garnet growth at  $1347 \pm 10$  Ma (garnet core, MSWD = 4.5) and  $1102 \pm 42$  (garnet rim, MSWD = 42).

In sample 51DTB19, concordant U–Pb monazite ages between ca. 1400 and 1200 Ma are significantly older than the existing Lu–Hf garnet ages of ca. 1081–1024 Ma (Zirakparvar et al., 2010). Seven concordant monazite inclusions in garnet yield a  $1318 \pm 14$  Ma concordia age (MSWD = 0.1), consistent with monazite (but not garnet) growth during the ca. 1380–1300 Ma East Kootenay tectono-thermal event. The

youngest three concordant matrix monazite yield a  $1021 \pm 20$  Ma concordia age that corresponds well with the Lu–Hf garnet age. Collectively, the data suggest this rock experienced lower-pressure conditions (under which garnet was not stable) during the East Kootenay event than during the Grenville age event.

The apatite U–Pb data from this sample yield an age of  $56 \pm 7$  Ma (MSWD = 32; Fig. 4.8). The high MSWD of the Discordia regression suggests this sample contains more than one apatite age-population. The  $\sim 0.48$   $^{207}\text{Pb}/^{206}\text{Pb}$  common-Pb intercept of this age is also much lower than predicted by common-Pb evolution models (Albarede and Juteau, 1984). Consequently, we suggest that apatite within sample 51DTB19 grew during partial breakdown (dissolution) of an older radiogenic Pb-bearing phase, likely a population of older Cryogenian apatite similar to the ca. 676 Ma apatite within nearby sample 49DTB19 (Fig. 4.9). The preservation of some likely Cryogenian apatite within this sample indicates that the Cordilleran processes this rocks experienced were within the U–Pb partial retention (or growth/dissolution) zone of apatite which is around 450–550 °C (Kirkland et al., 2018).

In the internal zone of the Clearwater complex, the Eocene  $54 \pm 2.7$  Ma (MSWD = 1.3) apatite U–Pb age from sample 52DTB19 (Fig. 4.9) records cooling and exhumation of the Clearwater complex, and is contemporaneous with existing ca. 64–55 Ma metamorphic zircon rims and ca. 53 to 47 Ma  $^{40}\text{Ar}/^{39}\text{Ar}$  mica ages (Doughty et al., 2007). These data indicate that rocks in the internal zone of the complex were buried deeper during Cordilleran processes than the external flanking zones (e.g. Lang and Rice, 1985; see Fig. 4.4). However, the internal zone does preserve a record of hydrothermal metamorphic zircon growth at ca. 1396–1350 Ma within this sample (Brennan et al., 2022). Apatite within sample 52DTB19 gave an upper  $^{207}\text{Pb}/^{206}\text{Pb}$  common Pb intercept of  $\sim 0.50$ , which again is much lower than predicted by common-Pb evolution models (Albarede and Juteau, 1984). Consequently, it is likely that the apatite in this sample also grew during the breakdown/dissolution of an older unknown radiogenic Pb-bearing phase perhaps also apatite, consistent with Cordilleran conditions that exceeded the U–Pb partial retention zone of apatite ( $\sim 550$  °C; Kirkland et al., 2018).

Along the eastern flanking region of the Clearwater Complex, U–Pb monazite ages of  $1337 \pm 30$  Ma and  $1119 \pm 18$  Ma are present in sample 59DTB19 (Fig. 4.8). The apatite U–Pb results record a  $58 \pm 9$  Ma (MSWD = 1.4) age (Fig. 4.9) that is also consistent with Cordilleran burial and Eocene core complex formation. No apatite

record of an older Proterozoic component present in rocks further to the west (in samples 49DTB19 and likely 51DTB19) was found. Interestingly, the  $58 \pm 9$  Ma U–Pb apatite age in sample 59DTB19 gives an upper  $^{207}\text{Pb}/^{206}\text{Pb}$  common Pb intercept of  $\sim 0.83$ , which is similar to values predicted by common-Pb evolution models (Albarede and Juteau, 1984). Consequently, the apatite in this sample does not record breakdown/dissolution of an older common-Pb bearing phases, perhaps indicating that the Cyrogenian apatite growth present along the western external zone did not occur in the eastern zone.

#### **4.5.2 Linking the timing and conditions of metamorphism**

Integrating existing garnet Lu–Hf (Zirakparvar et al., 2010; Neishem et al., 2012) with U–Pb monazite and apatite data and phase equilibrium modelling allows the timing and conditions of metamorphism within the Clearwater region to be constrained, particularly along the western external zone of the complex. Along the western edge of the complex, monazite U–Pb ages (samples 49, 50, and 51DTB19) indicate that the rocks were affected by high- $T$  processes during the older ca. 1380–1300 Ma East Kootenay tectono-thermal event. However, only one sample (50DTB19) records similar, ca. 1350 Ma, Lu–Hf garnet (core) ages (Neishem et al., 2012). While in these same three samples, younger ca. 1115–1065 Ma Grenville ages are ubiquitously recorded in monazite U–Pb and Lu–Hf garnet ages (Zirakparvar et al., 2010; Neishem et al., 2012). The presence of East Kootenay garnet ages in only one sample, while Grenville-age garnets are present in all samples, suggests that along the western, external, region of the Clearwater complex garnet did not grow in most rocks during the earlier Kootenay event, but was a common product of subsequent Grenville-age metamorphism. This relationship is consistent with higher pressures during the latter, Grenville-age event.

#### ***East Kootenay metamorphism***

Determining metamorphic conditions for the older ca. 1380–1300 Ma East Kootenay tectonothermal event is difficult due to extensive overprinting by Grenvillian metamorphism. Near the northern extent of the Belt Basin, within the Matthew Creek metamorphic zone and nearby St. Eugene deposit, ca. 1350 Ma metamorphism of lower Belt Supergroup strata is constrained to  $\sim 580$ – $650$  °C and 3.0–4.0 kbar (McFarlane and Pattison, 2000) and 490–510 °C and 3.6–4.0 kbar (Pattison and Seitz, 2012), consistent with ‘warm’ average geothermal gradients of

$\geq 35$  °C/km (e.g. Tucker et al., 2015). These relatively low-*P* conditions identified within the northern extent of the Belt Basin, are consistent with our inference of limited garnet growth within the western Clearwater complex. The relatively low pressures, and warm geotherms are more characteristic of Buchan rather than Barrovian metamorphism (e.g. Pattison and Spear, 2018; Pattison and Goldsmith, 2022 and references within). Consequently, the P-T conditions of the East Kootenay event are broadly consistent with sedimentary burial and heating due to coeval bimodal intrusions into the Belt Basin at this time (Doughty and Chamberlain, 1996).

The scattering of older ca. 1400–1200 Ma Lu–Hf garnet ages further east in the Clearwater complex (Neishem et al., 2012), suggest that this event may be better preserved, or that pressures of East Kootenay metamorphism were greater, further to the east. In sample 50DTB19, the thermodynamic modeling and mineral assemblages suggest that ca. 1350 Ma garnet core growth requires conditions of at least 620 °C and ~5 kbar. However, only a few kilometers in either direction, there is no evidence for garnet growth at this time. Thus, we speculate that the localized garnet growth within the western portion of the Clearwater complex at ca. 1350 Ma could reflect localized higher T/P conditions from documented nearby ca. 1380–1330 Ma magmatism within the Belt Basin (e.g. Doughty and Chamberlain, 1996; Lewis et al., 2007; Pearson and Link, 2021).

### ***Grenville-age metamorphism***

Rocks in the western portion of the Clearwater complex mostly preserve ca. 1115–1065 Ma metamorphic ages and Barrovian metamorphic assemblages that provide robust *P–T* constraints. Sample 49DTB19, which preserves ca. 1090–1070 Ma U–Pb monazite and ca. 1060 Ma Lu–Hf garnet ages (Fig. 4.8B), has an assemblage consistent with peak metamorphic conditions of 6.5–7.5 kbar and 640–654°C (Fig. 4.9). Sample 50DTB19 records ca. 1100 Ma garnet Lu–Hf rim ages and ca. 1060 Ma monazite U–Pb ages (Fig. 4.7B), and also records similar metamorphic conditions of ~6.8–7.8 kbar and ~650–670°C (Fig. 4.9). Sample 51DTB19, contains ca. 1080–1020 Ma Lu–Hf garnet ages, ca. 1020 Ma U–Pb monazite ages (Fig. 4.7B) and a mineral assemblage that suggests metamorphic conditions of ~620–680 °C and ~5.0–6.8 kbar (Fig. 4.10). Collectively, these values suggest that the rocks along the western region of the Clearwater complex were buried to a depth of ~17–22 km or greater (assuming isostatic pressures and an average crustal density of 3.0–2.7 g/cm<sup>3</sup>) around ca. 1080–

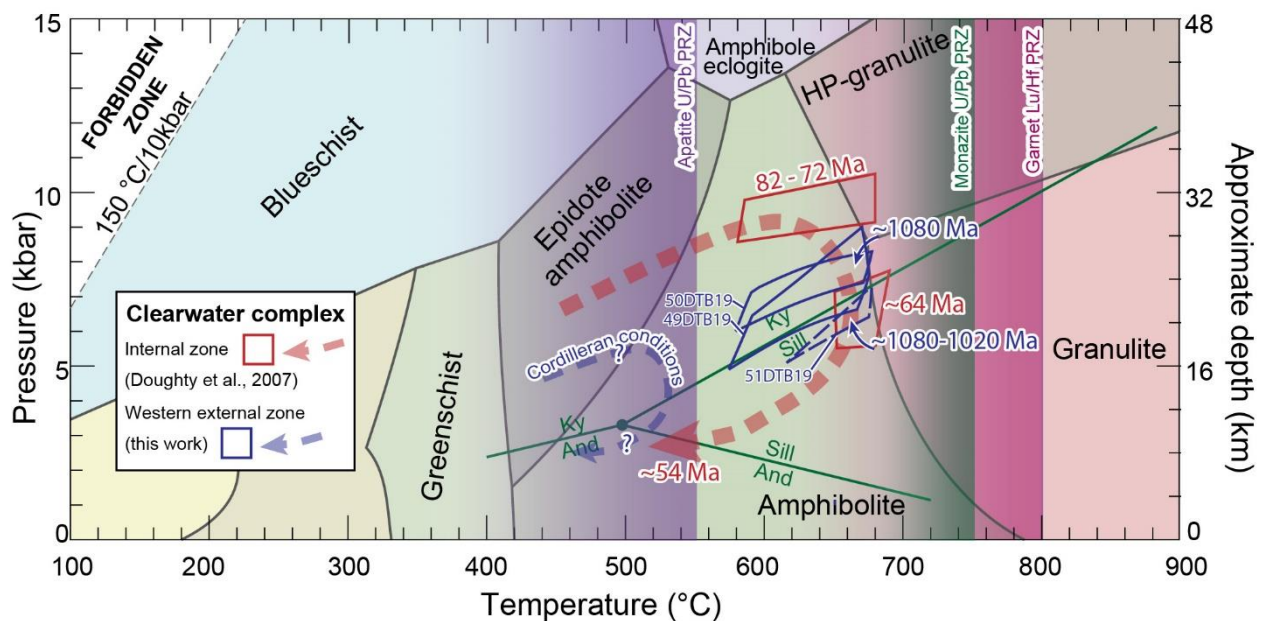
1020 Ma (Fig. 4.13). These results also imply a moderately-high average geothermal gradient ( $\sim 23\text{--}36^\circ\text{C}/\text{km}$ ; e.g. Tucker et al., 2015).

In the central portion of the Clearwater complex, Doughty et al. (2007) assign peak metamorphic conditions of  $\sim 9\text{--}10$  kbar and approximately  $650^\circ\text{C}$  at ca. 82–72 Ma, followed by a significant clockwise  $P\text{--}T$  path associated with core complex formation and exhumation from ca. 64 to 54 Ma (Fig. 4.13). These conditions suggest burial to a depth of  $\sim 31\text{--}38$  km, along a geothermal gradient notably cooler ( $17\text{--}21^\circ\text{C}/\text{km}$ ) than recorded by the rocks in the western external zone of the complex. The well-defined  $54 \pm 3$  Ma (MSWD = 1.3) and  $58 \pm 9$  Ma (MSWD = 1.4) apatite U–Pb ages reported here in the central and eastern portions (samples 52 and 59DTB19; Fig. 4.9) of the complex are in good agreement with this Cordilleran  $P\text{--}T\text{--}t$  pathway, and support Clearwater core complex exhumation in the late Paleocene to early Eocene (Fig. 4.13).

The  $676 \pm 13$  Ma (MSWD = 2.8) apatite U–Pb age within sample 49DTB19, and the over-dispersed  $56 \pm 7$  Ma (MSWD = 22) age recorded in sample 51DTB19 (Fig. 4.9), which we interpret to record partial resetting/recrystallization of an older Cryogenian apatite population, indicates that the western external region likely did not exceed  $\sim 550^\circ\text{C}$  (the apatite U/Pb partial retention zone) during Cordilleran orogenesis. As discussed, within these same samples the higher temperature Lu–Hf garnet or U–Pb monazite systems notably record no indication of this Cryogenian event, which is previously not identified regionally in other mid-temperature thermo-chronometers. Given that the Cryogenian apatite ages are the same as nearby (well-agreed upon) rift-related volcanism (the ca. 685–650 Ma Edwardsburg volcanics and Big Creek plutonic suite; Lund et al., 2003; 2010), which are not associated with any known super-greenschist facies metamorphism in this region, the Cryogenian apatite ages likely reflect a fluid-flow/thermal event associated Rodinian rifting. The absence of Cryogenian Lu–Hf garnet and U–Pb monazite ages indicate that this event did likely not result in re-equilibrium of the metamorphic assemblages from peak conditions at ca. 1115–1065 Ma (Fig. 4.13).

Interestingly, sample 49DTB19 contains the strongest signal of the ca. 1080 Ma Grenvillian event with complementary ca. 1090–1060 Ma U–Pb monazite and Lu–Hf garnet ages (Fig. 4.8), as well as the clearest ca. 676 Ma U–Pb apatite age of this Cryogenian rifting event (Fig. 4.9). Consequently, sample 49DTB19 is likely located proximal to a crustal/basin-scale structure that preferentially facilitated localized

recording of these tectono-thermal events. The presence of Cryogenian apatite populations (samples 49 and 51DTB19) and/or more radiogenic than expected common-Pb values suggesting breakdown of an older apatite population (sample 52DTB19) along the central and western region of the complex, while no indication of an Cryogenian apatite population (or breakdown of an older more radiogenic population) in the eastern region of the complex (59DTB19; see Fig. 4.4 and 4.9) could also perhaps reflect a Proterozoic structure (with a Cryogenian history) between these two regions.



**Fig. 4.13:** Pressure–temperature–time pseudosection interpretations for the internal zone of the Clearwater complex (Doughty et al., 2007) that reflect Cordilleran conditions from ca. 82–54 Ma, and the western external zone (this work) that reflect the ca. 1080–1020 Ma “Grenvillian” conditions recorded in the corresponding samples.

#### 4.5.3 A revised regional tectonic model

Rocks of the western Clearwater complex yield important new constraints that allow for a revised tectonic model for the Belt Basin region of western Laurentia during the Nuna to Rodinia transition. Insights into the absolute P/T conditions of ca. 1380–1300 Ma East Kootenay tectono-magmatic event in the studied region are limited due to overprinting by higher-grade ca. 1115–1065 Ma metamorphism. However, we note that the relatively extensive monazite growth, but only localized garnet crystallization identified during this event are consistent with high temperature, lower pressure, Buchan-style metamorphism similar to what workers (e.g. McFarlane and Pattison, 2000; Pattison and Seitz, 2012) identify coevally within similar rocks in



the northern portion of the Belt Basin, and attribute to extension and magmatism within the basin.

From the investigated samples, the peak metamorphic event along the western extent of the Clearwater complex give more robust insights into the ca. 1115–1065 Ma Grenvillian event within the Belt Basin. Tectonic models for this event within the Belt Basin region must now account for burial of Belt Supergroup rocks to a depth of ~17–22 km or greater along a moderately-high average geothermal gradient (~23–36°C/km) at ca. 1100–1065 Ma. The most recent model to explain this period of Barrovian-style Grenville-age metamorphism within the Clearwater complex advocates for continental collision and crustal thickening mechanisms at ca. 1090–1070 Ma based primarily on Lu–Hf garnet ages, and the generalized metamorphic grade/fabric suggested by the accompanying mineral assemblages (Neishem et al., 2012). However, recent documentation of metamorphism of pelitic rocks to conditions of ~680–660°C and 5.5–10.5 kbar, within a deep intraplate rift-basin (Tucker et al., 2015) caution against the ubiquitous attribution of compressional tectonic thickening to explain medium-pressure metamorphism unless supported by other lines of regional geologic evidence.

The Stenian Period (ca. 1200–1000 Ma) is admittedly predominantly a time of orogenesis and crustal thickening across southern Laurentia, associated with development of the Grenville province (e.g. Mulder et al., 2017). However, during this period of generally widespread crustal thickening, there is a distinctive period of continent-wide mantle/thermal upwelling, intraplate volcanism, reactivation of crustal weaknesses, and extension from ca. 1110–1040 Ma within Laurentia. Manifestations of this event are found in the 1110–1085 Ma Midcontinent Rift (e.g. Swanson-Hysell et al., 2018), 1115–1078 Ma Pikes Peak batholith (Guitreau et al., 2016), 1094–1080 Ma Southwestern Laurentian Large Igneous Province (Timmons et al., 2001; Bright et al., 2014), and 1090–1040 Ma Bylot and correlative basins in northern Laurentia (Greenman et al., 2021; see Fig. 4.2). However, although located <500 km from the Grenville orogenic front, this period of mantle upwelling and extension within the Midcontinent rift was punctuated by localized ca. 1090–1030 Ma (Cannon et al., 1993; Cannon, 1994), and ca. 1010–980 Ma; Swanson-Hysell et al., 2019; Hodgin et al., 2022) structural basin inversion, associated with the Ottawa and Rigolet phases of the Grenville orogeny. This punctuated inversion of the Midcontinent Rift indicates that Grenville orogenic deformation propagated far into the Laurentian continental

interior, often associated with inherited rift structures in regions weakened by recent thermal upwelling.

Consequently, the possibility of the ca. 1100–1065 Ma metamorphism within the Clearwater complex and Belt Basin to reflect similar processes (e.g. thermal upwelling and/or inversion of inherited structures) as is recognized coevally within the Midcontinent Rift is worth exploring. The work of various other geologists over the past couple of decades has demonstrated that significant syn-depositional, but commonly reactivated (with most reactivation generally attributed to Neoproterozoic rifting, and Mesozoic and younger Cordilleran deformation), crustal-scale faults break the Belt Basin up into segments with varying subsidence/exhumation histories (e.g. Winston, 1991; Sears, 2007; Lydon, 2010; Lonn et al., 2020 and references within). From north to south, the most significant of these crustal-scale structures consist of the St. Mary-Moyie and associated Rocky Mountain Trench fault-systems, the Showshoe fault and Jocko Line that border an intrabasin sydepositional high called the Central horst, and to the south, the Perry Line (Fig. 4.14A; Winston, 1986; Sears, 2007; Lydon, 2010). Some workers (e.g. Bookstrom, 2016) have suggested that the ca. 1380–1300 Ma East Kootenay and ca. 1115–1065 Ma Grenvillian events are generally pervasive along the western margin of the Belt Basin. However, alternatively we argue that it is only locally within the vicinity of these significant syn-depositional Belt Basin structures, that a clear record of ca. 1380–1330 Ma “East Kootenay” magmatism and metamorphism, and ca. 1115–1065 Ma “Grenville” metamorphism is identified.

In the northern region of the Belt Basin, adjacent to significant syndepositional structures of the northeast-southwest trending, St. Mary-Moyie Fault system, the Matthew Creek metamorphic zone and nearby world class Sullivan Pb-Zn-Ag deposit are present (Fig. 4.14A). Within the Matthew Creek metamorphic zone, sillimanite-grade amphibolite facies rock of the lower Belt Supergroup are of notably higher metamorphic grade than nearby greenschist facies Belt Basin rocks (McFarland and Pattison, 2000), and are intruded by the ca. 1365 Ma Hellroaring Creek and ca. 1335 Ma Matthew Creek stocks (Aleinikoff et al., 2015). These sillimanite-grade amphibolite facies rocks yield a spread of U–Pb monazite ages from ca. 1380–1000 Ma, with two distinct age-peaks at ca. 1350–1340 and 1080 Ma (Fig. 4.15; McFarlane, 2015). Associated with significant syn-depositional faults, cassiterite (tin-ore, SnO<sub>2</sub>) from the nearby Sullivan SedEx deposit yields U–Pb ages with a major ca. 1470 Ma age-peak associated with the timing of mineralization, and two significant overprinting

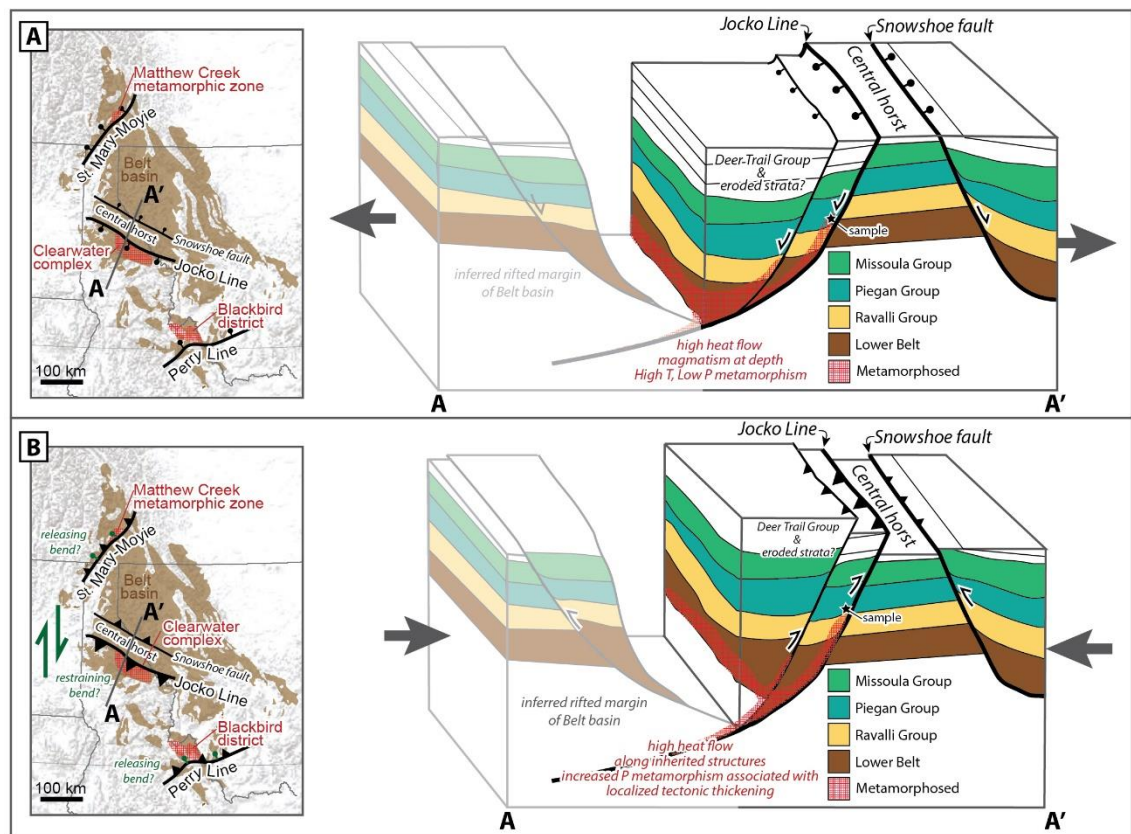
events at ca. 1380 and 1075 Ma that reflect dissolution-reprecipitation processes associated with localized thermal perturbations (Slack, 2020). Proterozoic structures of the St. Mary-Moyie Fault system also extend further southeast into the Priest River region, where rare ca. 1100 Ma metamorphic zircon rims are present (Doughty and Chamberlain, 2008).

The southernmost syn-depositional fault-system of the Belt Basin occurs along the east-west trending, down to the north, Perry Line (Fig. 4.14A; Winston, 1986; Link et al., 2007). The trend of the Perry Line extends west (e.g. Sims et al., 2004), into the Blackbird Co-Cu-Au district, where strata of the Lemhi Subgroup of the Belt Basin hosts Mesoproterozoic mineralization (Aleinikoff et al., 2012; Bookstrom et al., 2016). Ca. 1370 Ma diabase, diorite, and migmatites are also associated with the mineralized Belt Basin strata in the Blackbird district. Metamorphic barometry on these migmatites suggest initial pressures equivalent to ~14 km of burial, and subsequent pressure increases equivalent to ~20 km of burial (Doughty and Chamberlain, 2010). Monazite and xenotime U–Pb geochronology indicates that initial mineralization formed at ca. 1360 Ma, and experienced a subsequent Mesoproterozoic thermal event at ca. 1060 Ma (Fig. 4.15; Aleinikoff et al., 2012).

Within the central portion of the Belt Basin, the region investigated by this study, syndepositional basin-scale structures include a central horst block bounded to the north by the Snowshoe fault and to the south by the Jocko Line. Associated with the Snowshoe and Jocko syn-depositional structures are world-class Pb-Zn-Co-Ag deposits of the Coeur d’Alene and associated districts (e.g. Hobbs et al., 1965). The Jocko Line is of particular interest to this study, as it is a down to the south, significant syn-depositional Belt Basin fault that is an antecedent to the Lewis and Clark fault zone (Fig. 4.14; Winston, 1991; Sears, 2007). Notably, Lewis and Clark fault zone structures accommodated formation/exhumation of the Clearwater metamorphic complex during Cordilleran deformation (Doughty et al., 2007). As discussed, U–Pb monazite ages from the Clearwater complex contains distinct ca. 1380–1340 and 1080–1060 Ma bimodal age-peaks, that are similar to those found adjacent to other major syn-depositional Belt Basin structures (Fig. 4.15). Consequently, Grenville-age ca. 1080–1060 Ma metamorphism within the Belt Basin is associated with significant pre-existing crustal-scale structures. These structures were likely active in a normal-sense during the main ca. 1470–1380 Ma interval of

Belt Basin subsidence and were conduits for magmatism and mineralization during ca. 1370–1330 Ma East Kootenay metamorphism within the Belt Basin (Fig. 4.14A).

Importantly, the record of Grenville-age metamorphism is also associated with these major Belt Basin syn-depositional normal-fault structures. The sampled metapelitic units are from the Wallace Formation which occupies a middle-stratigraphic position of the Belt Supergroup, with an estimated ~9 km of underlying strata (Winston, 2016). The possibility that the sampled Wallace Formation is actually miscorrelated, lower Belt strata, is unlikely as the sampled unit contains consistent carbonate horizons absent in nearby lower Belt units, and also lacks the syn-depositional mafic sills common in nearby lower Belt strata (e.g. Lewis et al., 2001; 2005; 2007). Maximum thickness estimates for Belt Supergroup are generally ~20 km



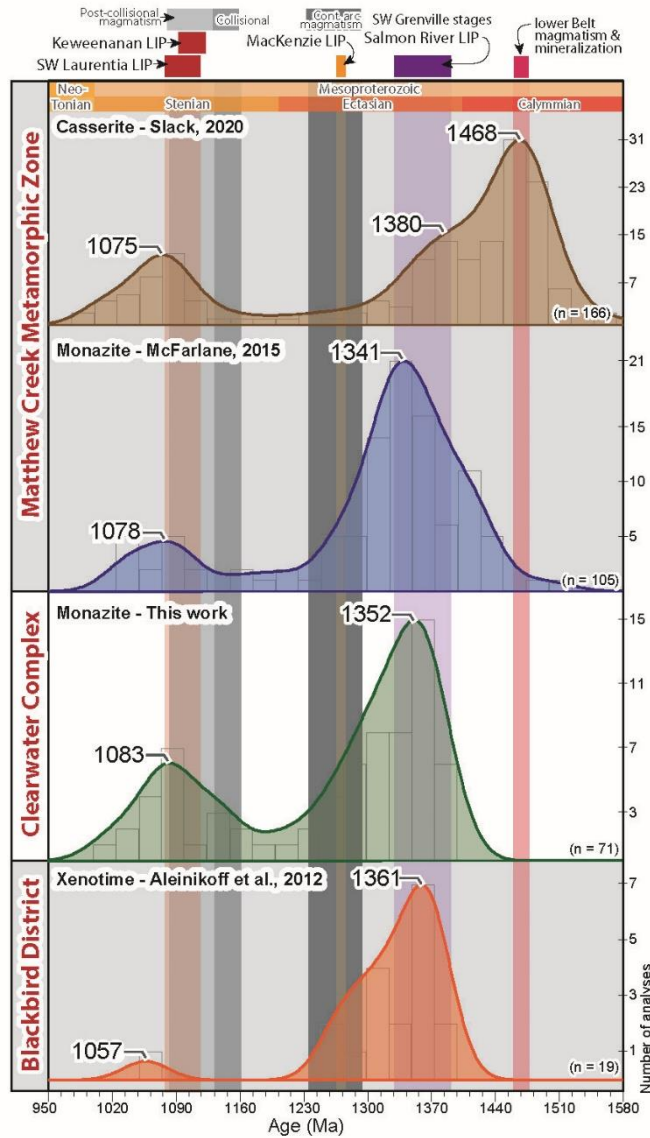
**Fig 4.14:** Generalized map (adapted from Sears, 2007 and Lonn et al., 2020), and schematic cross section cartoons of significant Belt basin syndepositional faults/fault zones at (A) the suggested occurrence of ca. 1380–1330 Ma East Kootenay, Buchan-style, metamorphism due to high heat flow, mineralization and magma intrusion into the Belt basin, and at (B) the suggested occurrence of ca. 1110–1060 Ma Grenville-age, Barrovian-style, metamorphism due to high heat flow and localized inversion of the Belt basin due to rejuvenation of inherited fault zones. Various mechanisms are worth investigating for this apparent Grenville-age tectonism, including a predominately compressional stress-field resulting in reverse-sense activation (shown by black faults) of inherited weaknesses, or an predominately transpressional stress-field resulting in some localized reverse sense, and perhaps some nominal or normal activation sense structures (alternative fault type shown in green) associated with restricting or releasing bends in the margin.

(Harrison, 1972) and re-correlation of the overlying Deer Trail Group strata (Box et al., 2020) may indicate at least another ~2.5 km (Miller and Whipple, 1989) of poorly preserved strata atop upper Belt strata. Consequently, at ca. 1080 Ma, the preserved stratigraphy can only account depositional burial of the Wallace Formation to depths of about 12–14 km, while the thermodynamic modelling results indicate pressures requiring at least 17–22 km of burial. It is possible that the stratigraphic thicknesses are underestimates due to subsequent erosion.

Nevertheless, the results presented in this work lend further support to earlier interpretations (Zirakparvar et al., 2010; Neishem et al., 2012) that Grenville-age tectonism in the Clearwater region of the Belt Basin requires a component of tectonic thickening or structural burial. Given the basin-wide association of this metamorphism with crustal-scale structures, and coeval timing to similar events elsewhere within Laurentia, we suggest this localized Grenville-age metamorphism within the Belt Basin was facilitated by enhanced heat flow and localized structural burial along inverted rift structures. This interpretation suggests that at ca. 1100–1020 Ma the Belt Basin experienced a period of mantle/thermal upwelling (e.g. Doughty and Chamberlain, 2008) and localized structural inversion (Fig. 4.14B), perhaps partially analogous/correlative to coeval processes recognized within the Midcontinent Rift and Southwest Laurentian large igneous province (e.g. Cannon, 1994; Bright et al., 2014; Swanson-Hysell, 2019; Hodgins et al., 2022). However, these processes in the Belt Basin differ notably from coeval Midcontinent and Southwest Laurentian processes. In particular, this time interval was anagmatic within the Belt Basin and there is no recognized associated sedimentation (Link et al., 1993; Brennan et al., 2021a; 2021b).

Identification of field-based relationships revealing this probable localized Grenville-age inversion of the western Belt Basin may be hampered by the fact that these same inherited structures were subsequently reactivated multiple times during Neoproterozoic rifting, and Mesozoic and younger Cordilleran orogenesis (e.g. Doughty et al., 2007; Lund et al., 2010; Bookstrom et al., 2016). During Neoproterozoic rifting the Belt Basin segment of the margin also experienced less subsidence and more exhumation than adjacent segments (Lund, 2008; Link et al., 2017; Brennan et al., 2020), which could have eroded a limited Grenville-age sedimentary record if it was ever present. However, as the geochronology and thermodynamic modeling results indicate, Grenville-age tectonism is indeed present

in western Laurentia. The substantiation of this finding has significant implications for global models of supercontinent Rodinia formation.



**Fig. 4.15:** Kernel Density U–Pb age spectra (plotted with 25 Myr band and binwidth) from amphibolite grade Belt Supergroup metapelites within the Matthew Creek metamorphic zone (Casserite, Slack, 2020; Monazite, McFarlane, 2015), the Clearwater complex (Monazite, this work), and the Blackbird district (Xenotime, Aleinikoff et al., 2012). The ages of corresponding southwestern Grenville stages (Mulder et al., 2017), lower Belt syndepositional mineralization, and notable Laurentian Large Igneous Provinces (LIPs; Ernst et al., 2008 and references within) are indicated. Note all three locations contain a bimodal age-distribution with 1380–1340, and 1080–1060 Ma age-peaks.

#### 4.5.4 Global Implication of Grenville-age Barrovian metamorphism within the Belt Basin

Most paleogeographic reconstructions favor the assembly of continents along western Laurentia at ca. 1100–900 Ma associated with the greater amalgamation of supercontinent Rodinia (e.g. Li et al., 2008; Mulder et al., 2018; Merdith et al., 2020). Consequently, the presented evidence for ca. 1110–1020 Ma Barrovian metamorphism within the Belt Basin presents several new constraints to refine these global models.

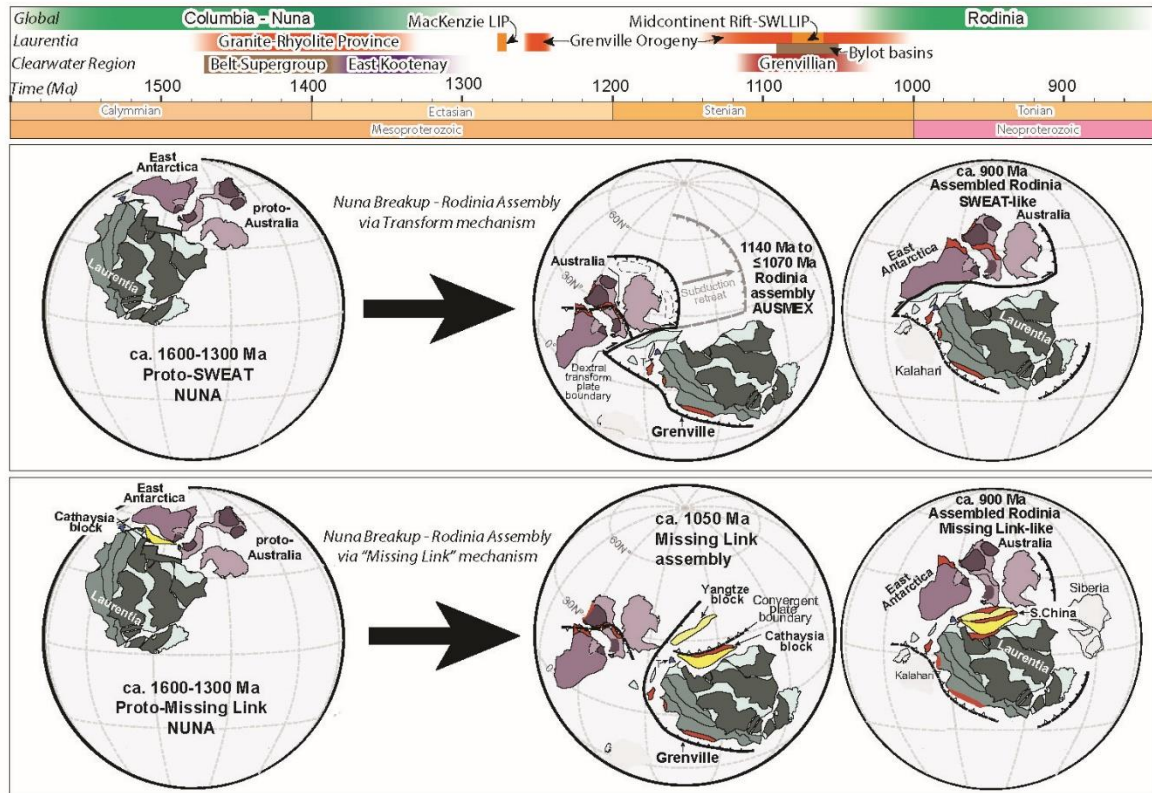
Firstly, these results suggest that the onset of amalgamation of continents west of the Belt Basin region initiated around 1110 Ma, and concluded by ca. 1020 Ma. Given the absolute absence of recognized tectonism in the Belt Basin region from ca. 1020–780 Ma (e.g. Brennan et al., 2021a), we suggest this ca. 1020–780 Ma interval was the stable duration of Rodinia in this region. Notably, this timeframe is longer than suggested by most leading global models (e.g. Li et al., 2008; Merdith et al., 2020) that instead suggest the main amalgamation of Australia, Antarctica, western Laurentia, and perhaps South China occurred primarily between ca. 1040–900 Ma, and began breaking up around ca. 830–820 Ma. This discrepancy could indicate that the Belt Basin region only records the initial onset of this assembly process and lacks a record of the initial onset of breakup processes, or perhaps the absolute timing of these global models require refinement.

Importantly, the Clearwater region is ~1800 km from the nearest Grenville-orogenic front in southwest Laurentia (see Fig. 4.2). This distance is over three times further than the ~500 km from the Grenville orogenic front in eastern Laurentia and the reactivated/inverted structures of the Midcontinent Rift. Thus, it is doubtful that Grenville orogenic stresses in southern and eastern Laurentia could solely result in the far-field inversion of the Belt Basin, especially since the record of structural inversion appears concentrated more in the western extent of the basin. Consequently, Rodinia models with continental blocks exhibiting tectonism adjacent to western Laurentia at ca. 1110–1020 Ma should be favored over those that lack a satisfactory mechanism to account for the Grenville-age metamorphic record of the Clearwater complex.

One such model that provides a satisfactory Grenville-age mechanism is the “Missing-Link” model (Li et al., 1995). In this model, final Rodinia assembly may have occurred along the ca. 1040–900 Ma Sibao orogen in South China, which is suggested to have been located between western Laurentia and Australia/Antarctica (Fig. 4.16). Additionally, the “Missing-Link” model also has several other strengths. Notably, South China also provides a satisfactory Neoproterozoic plume and rifting record that could account for discrepancies in corresponding records of western Laurentia and eastern Australia (e.g. Li et al., 2008). However, the absence of any non-Laurentian derived Grenville-age sedimentation in (or along) the Belt region (see Brennan et al., 2021a; 2021b), contradicts a long-held prediction of this model (Li et al., 1995; 2002; 2008; Yao et al., 2017). Others have also utilized paleomagnetic constraints (Park et al., 2021; Chang et al., 2022), and the presence of moderately



widespread ca. 950–830 Ma magmatism (Yao et al., 2019) in South China to instead argue that South China was located along the periphery, not center, of Rodinia. Based primarily on paleomagnetic data and broadly similar Neoproterozoic histories, alternative “Missing-Link” blocks with records of Grenville-age tectonism such as Tarim have also been proposed (Wen et al., 2017; 2018).



**Fig. 4.16:** Geologic timeline of the Nuna to Rodinia supercontinent transition, the relevant Laurentian-scale geologic events and local events within the Clearwater region. Below the timeline are two competing tectonic models for the Nuna to Rodinia transition along western Laurentia that could account for the ca. 1110–1020 Ma metamorphic record of the Clearwater complex. The upper model proposes a “SWEAT-like” (e.g. Moores, 1991) assembly of Rodinia along a large dextral transform boundary (after Mulder et al., 2018; Meredith et al., 2020). The lower “Missing-Link like” model proposes that the assembly of Rodinia occurred along the intervening Cathaysia and Yangtze blocks which contain a ca. 1100–900 Ma collisional (Sibao) orogen (after Li et al., 1995; 2008; 2013; Yao et al., 2017).



Alternative models besides the “Missing-Link” also warrant consideration. The most notable of which are those that advocate for the final assembly of western Laurentia and Australia/Antarctica along a dextral transpressional/transtensional boundary perhaps better recorded in smaller intervening blocks, such as Tasmania and the Western South Tasman Rise (Fig. 4.16). A strength of these models, is that Tasmania and the Western South Tasman Rise contain ca. 1300–1080 Ma metamorphism and provenance ties to western Laurentia (Halpin et al., 2014; Mulder et al., 2018; Merdith et al., 2020; Brown et al., 2022). It is also worth noting, that some alternative Missing-Link models, such as those that favor Tarim between western Laurentia and Australia/Antarctica (Wen et al., 2018) also advocate for dextral transpressional assembly of Rodinia along the western Laurentian margin.

In these models of transpressional Rodinia assembly, the observed ~6 km discrepancy between calculated burial depths and estimates for the thickness of overlying strata in the Clearwater region at ca. 1080 Ma, suggest that the Clearwater region (and the Jocko Line; see Fig. 4.14) could have acted as a localized restraining bend boundary along the western edge of the Belt Basin during this period (Fig. 4.14B; 4.16), consistent with the proposed dextral motion. If correct, based on the orientations of the Perry Line, and St. Mary-Moyie fault zones, the dextral assembly models predict these regions would have been predominantly releasing bends along the margin. This relationship may suggest nominal Grenville-age inversion and limited related Barrovian-style metamorphism, but perhaps still a record of thermal upwelling and/or subsidence, in these areas. Consequently, dextral transpressional models could explain why the ca. 1380–1330 Ma East Kooteney metamorphism along the Perry Line and St. Mary-Moyie Fault zone regions (Matthew Creek metamorphic zone and Blackbird district) appears to have a weaker Grenville-age overprint (e.g. Pattison and Seitz, 2012; Bookstrom et al., 2016) compared to the Clearwater complex.

## **4.6 Conclusions**

A key record to solving debates on the style and mechanism of the Nuna to Rodinia transition lies in the ca. 1100 Ma Grenville-age tectonic record of western Laurentia. Within the Clearwater complex of the Belt Basin in western Laurentia, existing Lu–Hf garnet geochronology (Zirakparvar et al., 2010; Neishem et al., 2012) and new monazite and apatite U–Pb results and Thermocalc (pressure-temperature) constraints from ca. 1450 Ma Belt Supergroup metapelitic strata record ca. 1350 and

1080 Ma metamorphism, and thus may contain a rare record of this supercontinent transition. However, variable overprinting by younger Mesozoic and Cordilleran orogenic processes has previously hindered interpretation of these older events. In the western external region of the Clearwater complex, metapelitic strata retain, or partially retain, Cryogenian apatite U–Pb ages associated with nearby rift-related magmatism during Rodinia breakup. These ages indicate that Cordilleran conditions did not exceed ~450–550°C (the U/Pb closure temperature of apatite) in this region. Consequently, the mineral assemblages within these rocks likely reflect an upper amphibolite grade, Barrovian-style, metamorphic event recorded by ca. 1100–1070 Ma garnet Lu–Hf and monazite U–Pb ages. Thermodynamic modeling of these metamorphic assemblages indicate they are consistent with pressures of 5–7.5 kbar and temperatures of 620–670°C, suggesting burial to a depth of ~17–22 km or greater around along a moderately-high average geothermal gradient of ~23–36°C/km.

These ca. 1100–1070 Ma metamorphic conditions are difficult to satisfy with static heating within the Belt Basin alone. Thus, these conditions advocate that thermal upwelling/heating and western tectonism facilitated rejuvenation of crustal-scale rift-structures which resulted in localized Grenville-age structural inversion, and crustal thickening within the Belt Basin. Manifestations of this critical period of thermal upwelling/heating, rejuvenation of crustal-scale structures, and sometimes localized structural rift inversion associated with Rodinia assembly may be recognized Laurentian-wide such as within the 1110–1085 Ma Midcontinent Rift (e.g. Swanson-Hysell et al., 2019), 1115–1078 Ma Pikes Peak batholith (Guitreau et al., 2016), 1094–1080 Ma Southwestern Laurentian Large Igneous Province (Timmons et al., 2001; Bright et al., 2014), and perhaps the 1090–1040 Ma Bylot and correlative basins in northern Laurentia (Greenman et al., 2021). Consequently, global geologic models must account for ca. 1100–1020 Ma Grenville-age tectonism in blocks adjacent to the western margin of Laurentia within supercontinent Rodinia. Future work should evaluate the viability of Grenville-age collisional or transpressional/transensional settings along the western margin of Laurentia and alleged conjugates such as Australia/Antarctica, South China and/or Tarim.

## 4.7. References

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## **Chapter 5: Recalibrating Rodinian rifting in the northwestern United States**

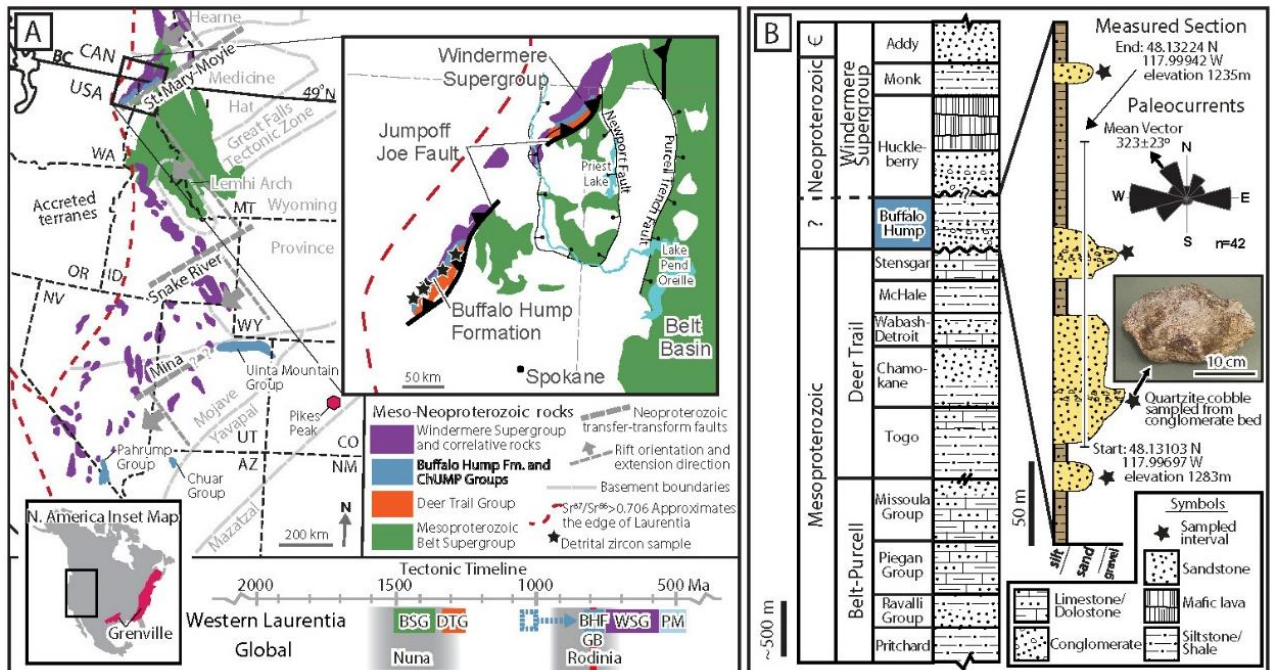
### **Abstract**

A lack of precise age constraints for Neoproterozoic strata in the northwestern United States (Washington State), including the Buffalo Hump Formation (BHF), has resulted in conflicting interpretations of Rodinia amalgamation and breakup processes. Previous detrital zircon (DZ) studies identified a youngest ca. 1.1 Ga DZ age population in the BHF, interpreted to reflect mostly first-cycle sourcing of unidentified but proximal magmatic rocks intruded during the amalgamation of Rodinia at ca. 1.0 Ga. Alternatively, the ca. 1.1 Ga DZ population has been suggested to represent a distal source with deposition occurring during the early phases of Rodinia rifting, more than 250 m.y. after zircon crystallization. We combined conventional laser-ablation split-stream analyses of U-Pb/Lu-Hf isotopes in zircon with a method of rapid (8 s per spot) U-Pb analysis to evaluate these opposing models. Our study of ~2000 DZ grains from the BHF identified for the first time a minor (~1%) yet significant ca. 760 Ma population, which constrains the maximum depositional age. This new geochronology implies that the BHF records early rift deposition during the breakup of Rodinia and correlates with sedimentary rocks found in other late Tonian basins of southwestern Laurentia.



## 5.1. Introduction

Correlation of pre-rift and rift strata across purportedly paired margins, such as Neoproterozoic strata of the Adelaide rift complex (southeastern Australia) and the Windermere Supergroup (western Laurentia), has long been used to define key piercing points for reconstructing supercontinents (e.g., Bell and Jefferson, 1987; Dalziel, 1991; Brookfield, 1993). In the northwestern United States, strata of the ca. 1.47–1.38 Ga Belt Supergroup, the ca. 1.3 Ga Deer Trail Group, the <1.1 Ga Buffalo Hump Formation (BHF), and the ca. <720 Ma Windermere Supergroup provide key constraints in the reconstructions of the supercontinents Nuna (ca. 1.6–1.3 Ga) and Rodinia (ca. 900–700 Ma; Fig. 5.1A; Ross et al., 1992; Ross and Villeneuve, 2003; Li et al., 2008; Kirscher et al., 2020). However, uncertain provenance and depositional ages of these key strata have resulted in varying proposed configurations, timings, and identities of alleged Laurentian conjugates such as Australia, Antarctica, and/or South China during the Proterozoic (e.g., Li et al., 2008; Box et al., 2020).



**Fig 5.1:** (A) Geologic map and time line of the North American Cordillera, and the northwestern United States study area, where Proterozoic rocks of the Belt Supergroup, Deer Trail Group, Buffalo Hump Formation, and Windermere Supergroup outcrop adjacent to the edge of Laurentia. Extent of the Grenville Province is shown on the inset map. Neoproterozoic rift geometry and structures are indicated (after Lund et al., 2010). (B) Stratigraphy of the study region (after Box et al., 2020) and photo of the cobble analyzed. Paleocurrent measurements indicate a mean vector to the northwest, roughly perpendicular to the Jumpoff Joe thrust fault. Abbreviations: ChUMP—Chuar–Uinta Mountain–middle Pahrump Group; BSG—Belt Supergroup, DTG—Deer Trail Group, BHF—Buffalo Hump Formation; GB—Gun Barrel, WSG—Windermere Supergroup, PM—passive margin; C—Cambrian. State/province abbreviations: BC—British Columbia, Canada; WA—Washington; MT—Montana; OR—Oregon; ID—Idaho; NV—Nevada; WY—Wyoming; UT—Utah; CO—Colorado; AZ—Arizona; NM—New Mexico

Approximately 10 km west of the westernmost Belt Supergroup in northeastern Washington State, coarse-grained quartzite and locally feldspathic siltite of the BHF and finer-grained argillite and carbonate rocks of the underlying Deer Trail Group crop out. An initial study suggested these rocks comprise a relatively continuous sequence that was correlative to the Piegan and Missoula Groups of the Belt Supergroup (Miller and Whipple, 1989). Ross et al. (1992) dated six detrital zircon (DZ) grains from the BHF and found four ca. 1.2–1.0 Ga grains and two older ca. 1.8 Ga grains. Because the BHF is coarser than Belt Supergroup rocks to the east, Ross et al. (1992) advocated a western source of ca. 1.1 Ga detritus into the BHF basin. More recent geochronology and field work have demonstrated that the Belt Supergroup and Deer Trail Group are characterized by DZ populations ca. 1.3 Ga and older, signifying deposition before the onset of the Grenville orogeny and Rodinia amalgamation (Ross and Villeneuve, 2003; Box et al., 2020). West of the Mesozoic Jumpoff Joe thrust fault (Fig. 5.1A), Deer Trail Group rocks are unconformably overlain by BHF strata containing abundant ca. 1.1 Ga (Grenvillian) DZs (Box et al., 2020).

Despite these developments, the original interpretation of a western ca. 1.1 Ga DZ source for the BHF (Ross et al., 1992) remained an important constraint for the reconstruction of Rodinia during its assembly (e.g., Li et al., 2008; Yao et al., 2017). However, others have argued that the youngest ca. 1.1 Ga DZ population does not represent the age of deposition, and instead the BHF may be the basal unit of the overlying Windermere Supergroup that constrains early Rodinia breakup (Box et al., 2020).

We adopted an integrated approach combining a rapid laser-ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) U-Pb technique with standard split-stream ICP-MS U-Pb and Lu-Hf analyses of DZ from the BHF. Our approach identified a minor and previously unrecognized DZ age population that permits recalibration of a key piercing point for Rodinia reconstructions.

## **5.2. Geologic setting and methodology**

The >350-m-thick BHF is characterized by alternating mudstone and sandstone strata that experienced regional lower-greenschist-facies metamorphism. Locally, the sandstone intervals include monolithic quartzite conglomerate beds containing cobbles >20 cm in diameter. The sandstone intervals decrease in thickness

and grain size up section, recording an overall fining-upward trend at the formation scale (Fig. 1B).

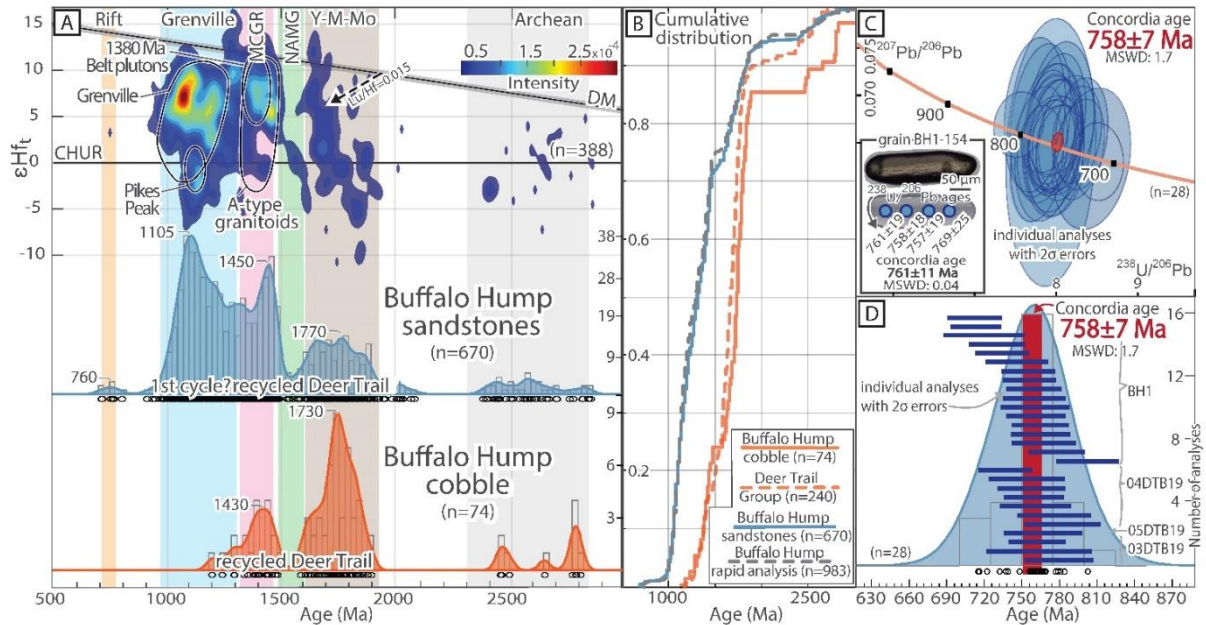
To reassess the depositional age and evaluate conflicting sources for the BHF detritus, DZ grains separated from four sandstone intervals and one quartzite cobble were mounted and imaged. Approximately 120 grains from each sample were analyzed by conventional laser-ablation split-stream ICP-MS (LA-SS-ICP-MS) methods for U-Pb and Lu-Hf isotopes at the GeoHistory Facility in the John de Laeter Centre at Curtin University (Perth, Australia) using a 38  $\mu\text{m}$  spot. This approach identified a single concordant  $756 \pm 30$  Ma ( $\epsilon\text{Hf}_t = 3.4 \pm 1.5$ ; all uncertainties reported at  $2\sigma$ ) zircon grain. An additional U-Pb session ( $\sim 400$  unknowns), utilizing a 20  $\mu\text{m}$  spot, identified two additional ca. 760 Ma grains. To further characterize this age, the remaining DZ grains from all four sandstone sample mounts ( $\sim 1200$  unknowns) were analyzed using a technique involving rapid (8 s each) LA-ICP-MS analysis. This analysis rate was achieved by bypassing the aerosol homogenization device tubing, omitting “cleaning pulses,” and reducing ablation and washout times (Chew et al., 2019). Rapid analysis identified additional ca. 760 Ma DZ grains in all four samples. Reanalysis of these grains via conventional U-Pb LA-ICP-MS methods provided robust age constraints. Age-population plots and maximum depositional age calculations only include zircon analyses with discordance less than 10% (Fig. 5.2).

### **5.3. Detrital zircon U-Pb and Lu-Hf isotope data**

The U-Pb/Lu-Hf LA-SS-ICP-MS analysis of DZ grains from all four BHF sandstone samples showed older components at ca. 2.8–2.4 Ga and ca. 1.9–1.6 Ga, along with two broad younger age peaks at ca. 1450 Ma and 1105 Ma that had mostly juvenile Lu-Hf ( $\epsilon\text{Hf}_t = +5$  to  $+10$ ) values (Fig. 2). Laurentian sources such as the Grenville Province (Fig. 1A, inset), the midcontinent (A-type) Granite-Rhyolite Province, the Yavapai-Mazatzal-Mojave Province, and Archean provinces to the east/southeast (e.g., Wyoming craton) are consistent with these age components (Whitmeyer and Karlstrom, 2007).

A juvenile ( $\epsilon\text{Hf}_t = +5$  to  $+10$ ) ca. 1380 Ma component may represent detritus derived from plutonic rocks that intrude the Belt Supergroup (Brennan et al., 2020). A minor ca. 1.1 Ga component with an evolved ( $\epsilon\text{Hf}_t = 0$  to  $-5$ ) signature indicates an additional source inconsistent with juvenile Grenville sources in southeastern Laurentia (Howard et al., 2015; Thomas et al., 2017). The 1.07 Ga Pike’s Peak

batholith of central Colorado is of similar age and isotopic composition ( $\epsilon_{\text{Hf}}$  of 2 to  $-3$ ; Howard et al., 2015) and was also undergoing rapid exhumation during the late Tonian (Flowers et al., 2020), indicating it could be a first-cycle source for this DZ component.



**Fig 5.2:** (A) Detrital zircon (DZ) U-Pb/Lu-Hf results for the Buffalo Hump Formation (BHF; Washington State, USA) sandstones and the BHF cobble. Age spectra are kernel density estimates (KDEs) with 25 m.y. bandwidths, which are also used for the histograms, and exclude repeated measurements on the targeted young grains. Lu-Hf data shown in the bivariate KDE plot also have 25 m.y. and 1  $\epsilon_{\text{Hf}}$  bandwidths. Average crustal evolution ( $\text{Lu}/\text{Hf} = 0.015$ ) is indicated. The  $\epsilon_{\text{Hf}}$  and age fields are from Thomas et al. (2017) and Brennan et al. (2020). CHUR—chondritic uniform reservoir; DM—depleted mantle; MCGR—Midcontinent Granite-Rhyolite; NAMG—North American magmatic gap; Y-M-Mo—Yavapai-Mazatzal-Mojave. (B) Cumulative distribution function highlights the similarity between the BHF cobble sample (orange) and the Deer Trail Group (dashed orange; Box et al., 2020), and BHF conventional sandstone (blue) and rapid sandstone (dashed gray) analyses (see the Supplemental Material). (C) Terra-Wasserberg concordia plot of ca. 760 Ma DZ grains, which yielded a Concordia age of  $758 \pm 7$  Ma ( $2\sigma$ ). Images (cathodoluminescence and reflected light) of grain BH1-154 in inset. Four analyses of this grain yielded similar ( $^{238}\text{U}/^{206}\text{Pb}$ ) ages and a Concordia age of  $760 \pm 12$  Ma ( $2\sigma$ ), indicating this grain experienced minimal Pb loss. MSWD—mean square weighted deviation. (D) KDE plot of young analyses with individual analysis bars ( $2\sigma$ ) and preferred  $758 \pm 7$  Ma maximum depositional age shown.

Twenty-eight (28) conventional U-Pb spot analyses on 21 ca. 760 Ma grains yielded a Concordia age of  $758 \pm 7$  Ma (mean square weighted deviation [MSWD] = 1.7, all given as a measure of concordance plus equivalence; Figs. 5.2C and 5.2D). All analyses overlapped with concordia at  $2\sigma$  and yielded alpha-doses and calculated densities (conservatively assuming damage accumulation initiated at crystallization) indicative of closed-system behavior of U-Pb (Murakami et al., 1991). Multiple robust analyses of single DZ grains allows better evaluation of potential Pb loss. Five individual zircon grains were large enough for multiple robust analyses. Four had similar concordia ages of  $762 \pm 10$  Ma ( $n = 4$ ; MSWD = 0.25; Fig. 2C),  $762 \pm 17$  Ma

( $n = 2$ , MSWD = 0.16),  $766 \pm 16$  Ma ( $n = 2$ , MSWD = 0.19), and  $770 \pm 12$  Ma ( $n = 3$ , MSWD = 0.34). The fifth multi-analysis grain gave a Concordia age of  $736 \pm 14$  Ma ( $n = 2$ , MSWD = 0.28).

The DZ ages from the quartzite cobble from the BHF conglomerate contain ca. 1.7 Ga and ca. 1.4 Ga components, with no grains younger than ca. 1.2 Ga (Figs. 2A and 2B).

## **5.4. Recalibrating rifting of Rodinia in western Laurentia**

### **5.4.1 New constraints from the Buffalo Hump Formation**

Based on the Concordia age of the young (ca. 760 Ma,  $n = 28$ ) analyses, we favor a  $758 \pm 7$  Ma (MSWD = 1.7) maximum depositional age for the BHF. This age is a purely descriptive measure of the youngest ages present (Figs. 2C and 2D). The single  $736 \pm 14$  Ma multi-analysis grain was found in the stratigraphically highest sampled interval (BH1) and may support a maximum depositional age that decreases up section from ca. 760 to 735 Ma, and a continued connection to the source of these ca. 760 Ma grains throughout BHF deposition. While ca. 760 Ma magmatic sources were present in eastern Laurentia (e.g., MacLennan et al., 2020), similar-aged source rocks are unknown from western Laurentia. Non-Laurentian source rocks with ca. 760 Ma ages occur in Tasmania ( $760 \pm 12$  Ma; Turner et al., 1998), Western Australia ( $755 \pm 3$  Ma; Wingate and Giddings, 2000), and the Yangtze block of South China ( $768 \pm 7$  Ma to  $751 \pm 10$  Ma; Li et al., 2003).

The large size of the BHF cobble (Fig. 1B) and the ages of the DZ it contains (Fig. 2A) suggest it was sourced from the underlying Deer Trail Group. To the east of the Mesozoic Jumpoff Joe thrust fault, BHF rocks are absent, and Deer Trail Group strata are ~3 km thinner than they are west of the fault (Box et al., 2020). However, Cambrian rocks are continuous across the fault (Lindsey and Gaylord, 1992). These relationships suggest that the Mesozoic Jumpoff Joe thrust fault may have been an ancestral down-to-the-northwest Neoproterozoic normal fault during BHF deposition. The fault is parallel to the Paleoproterozoic boundary between the Archean Hearne and Medicine Hat blocks (Fig. 1A), and likely exhumed Deer Trail Group rocks on the footwall to its east, interpreted as a rift shoulder, providing a proximal source for the cobbles that were recycled into the adjacent hanging-wall BHF rift basin.

### 5.4.2 A “ChUMP-B” Correlation and Implications

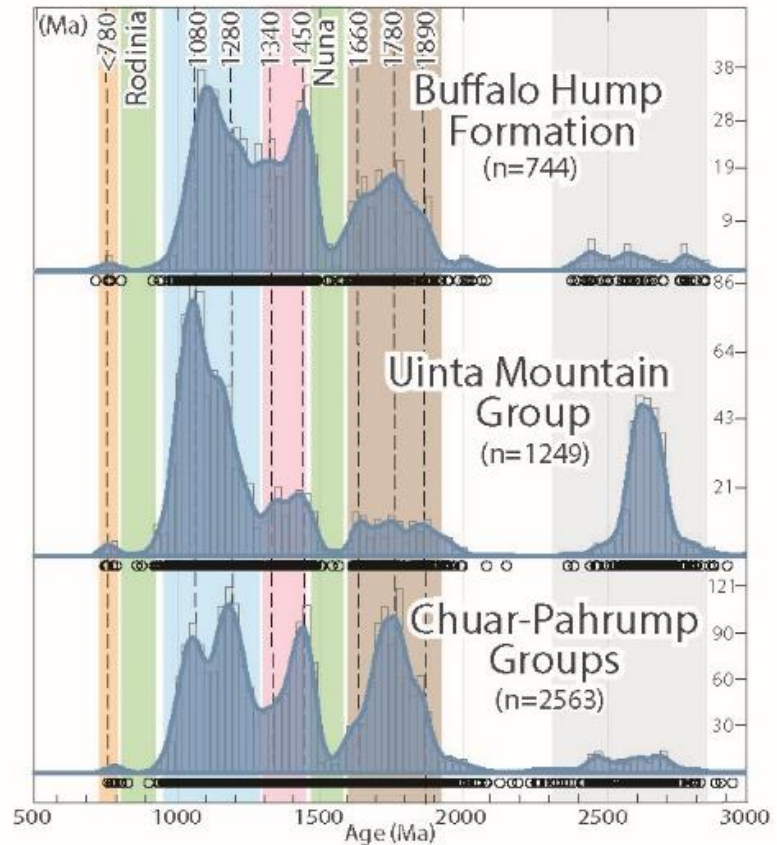
Detrital zircon grains with ca. 760 Ma ages are rare in western Laurentia and only known to occur in the Uinta Mountain Group (Fig. 1A). The Uinta Mountain Group has the same maximum depositional age ( $766 \pm 5$  Ma; Dehler et al., 2010) as the BHF and is also unconformably overlain by younger than 720 Ma Windermere Supergroup strata. A south-dipping, syndepositional fault is inferred to have bounded the Uinta Mountain Group basin to the north. The fault is parallel to, and likely related to, the east-west-trending Paleoproterozoic suture zone along the southern margin of the Archean Wyoming craton (Fig. 1A; Dehler et al., 2010). The orientations of the BHF and Uinta Mountain Group syndepositional faults suggest north-south to northwest-southeast extension along preexisting boundaries.

The Uinta Mountain Group is correlative to the Chuar and middle Pahrump Groups in southwestern Laurentia (Fig. 5.1A). These three coeval (Chuar–Uinta Mountain–middle Pahrump) groups comprise the “ChUMP” correlation (Dehler et al., 2017, and references therein). Similar to the BHF, ChUMP strata contain significant ca. 1.90–1.65 Ga, ca. 1.45 Ga, and ca. 1.3–1.0 Ga DZ age components (Fig. 3). A few late Tonian (ca. 800–720 Ma) DZ grains in ChUMP rocks constitute a comparable (<1.5%) component of the total DZ population. Basal rocks of the more southern Chuar and middle Pahrump Groups have ca. 780 Ma maximum depositional ages ( $<782 \pm 7$  Ma and  $<775 \pm 18$  Ma, respectively; Dehler et al., 2017; Mahon et al., 2014). Upper Chuar Group strata contain a  $742 \pm 6$  Ma interbedded tuff and are unconformably overlain by Cambrian rocks (Karlstrom et al., 2000, 2018). Middle Pahrump Group strata are overlain by younger than 720 Ma Sturtian diamictites (Mahon et al., 2014; Dehler et al., 2017).

ChUMP strata record an interval of intracratonic rift-basin formation that occurred after intrusion of the ca. 780 Ma Gunbarrel dike swarms (Harlan et al., 2003) but prior to widespread Laurentian rifting at ca. 720 Ma (Yonkee et al., 2014). The stratigraphic position, DZ ages, and tectonic setting of the BHF indicate that it is likely a correlative to ChUMP group strata in an expanded “ChUMP-B” (Chuar, Uinta Mountain, middle Pahrump Groups–Buffalo Hump Formation) correlation. The absence of ca. 900–800 Ma DZ grains in ChUMP-B strata (Fig. 5.3) likely reflects tectono-magmatic quiescence in western Laurentia during that interval due to its central location within Rodinia (e.g., Li et al., 2008).



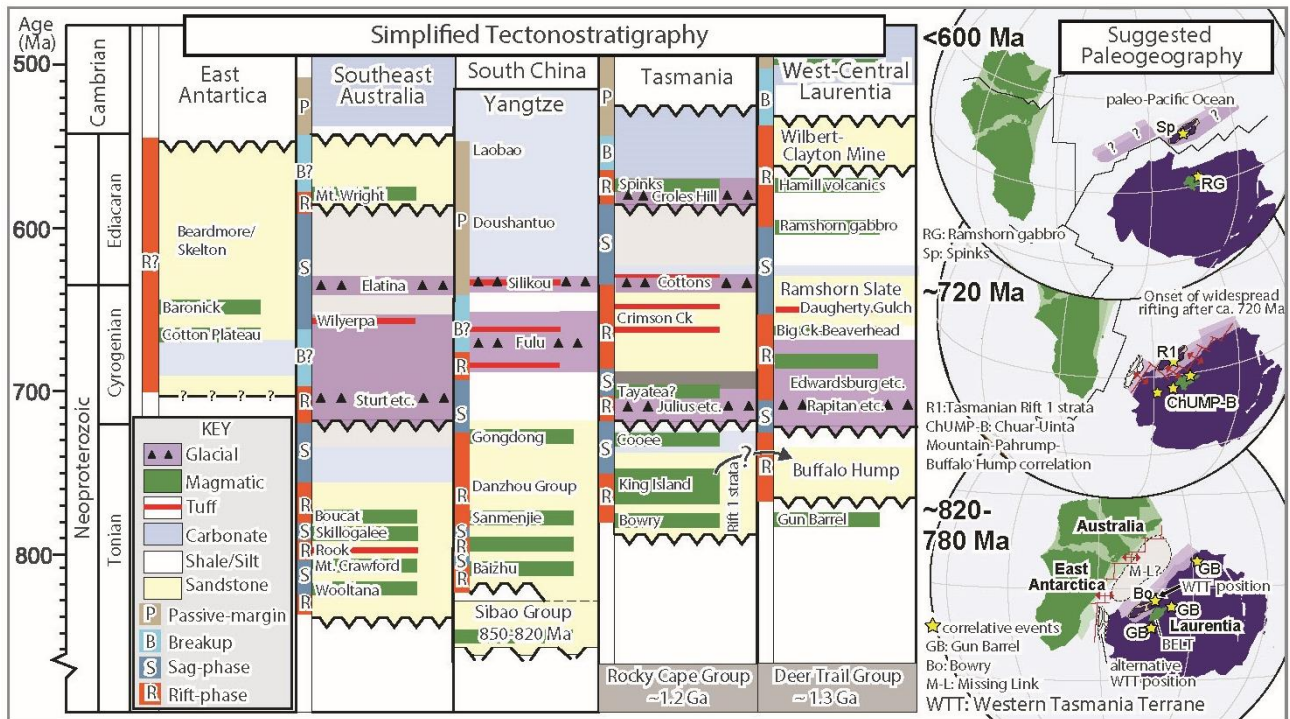
**Fig. 5.3:** Normalized U-Pb detrital zircon kernel density estimate (KDE) plots and main age components for late Tonian strata in western Laurentia.



The proposed ChUMP-B correlation extends the <780 to ca. 720 Ma interval of rift-basin development northward to the Belt segment (central Idaho to southeastern British Columbia) of the Cordilleran margin. In this region, intermittent Cryogenian–Cambrian Windermere Supergroup deposition and magmatism record predominately northeast-southwest extension, recurrent margin-parallel exhumation (e.g., the Lemhi arch; Fig. 1A), and accompanying breakup in the late Ediacaran–early Cambrian (Lund et al., 2010; Lund and Cheney, 2016; Brennan et al., 2020). Thus, it appears that the orientations of the BHF and Uinta Mountain Group syndepositional faults require earlier extension oblique to the main extension direction during deposition of the Windermere Supergroup. Southwestern Laurentia also likely experienced similar complex multiphase extension during this time (Timmons et al., 2001).

This history of rifting is difficult to reconcile with a single event between southeastern Australia (the Gawler craton) and western Laurentia (e.g., Yonkee et al., 2014). Southeastern Australia contains ca. 830–790 Ma synrift magmatism (Preiss, 2000), of which there is no record in the Belt region. Additionally, high-quality paleomagnetic constraints indicate Australia was widely separated or significantly offset from western Laurentia at ca. 750 Ma (Eyster et al., 2020). Thus, our proposed ChUMP-B correlation and the younger Windermere Supergroup strata support a

scenario with multiple rifting events west of Laurentia during Rodinia breakup (see Fig. 4). Intervening blocks, likely including western Tasmania, occupied the region between western Laurentia and Australia within Rodinia.



**Fig 5.4:** Simplified Neoproterozoic tectono-stratigraphy (and inferred rift/faulting, sag/thermal subsidence, lithospheric breakup, and passive margin phases) for west-central Laurentia (Belt segment) and suggested conjugate margins, and proposed paleogeographic cartoon (adapted from Preiss, 2000; Li et al., 2003; Mulder et al., 2020; Brennan et al., 2020, and references therein). In addition to being a magmatic source of ca. 780–720 Ma zircon, western Tasmania also contains a Neoproterozoic history (Mulder et al., 2020) similar to the Belt region (Lund et al., 2010; Brennan et al., 2020). This correlation is consistent with a multiple-rift scenario. Chuar, Uinta Mountain, middle Pahrupp Groups–Buffalo Hump Formation (ChUMP-B) strata may have formed in localized inboard rift basins generated from the breakup of Australia–Antarctica and Laurentia. Younger Windermere Supergroup records the rifting of additional intervening blocks, likely including western Tasmania, into an already open paleo–Pacific Ocean.

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# **Chapter 6: A tectonic model for the Transcontinental Arch: Progressive migration of a Laurentian drainage divide during the Neoproterozoic–Cambrian Sauk Transgression**

## **Abstract**

A widespread provenance shift recorded by passive margin strata of western Laurentia, from predominant Stenian (1.2–1.0 Ga) detrital zircon age components to their absence, occurred during the Neoproterozoic–Cambrian Sauk transgression and is commonly used as a ca. 540 Ma chronostratigraphic marker throughout the west/south-western United States. However, in Neoproterozoic–Cambrian strata of this region, we identify a probable shift from distal to more proximal Stenian-age zircon sources before a diachronous loss of Stenian detrital zircon age components. We suggest these provenance patterns reflect progressive subsidence of the passive margins surrounding Laurentia and concomitant relative uplift of the Transcontinental Arch, a broad and segmented northeast–southwest trending topographic high across the Laurentian midcontinent possibly due to lithospheric flexure. The Transcontinental Arch segments align with transverse rift structures of the Neoproterozoic–Cambrian Iapetan margin and the Mesoproterozoic Midcontinent Rift, perhaps reflecting rejuvenation of midcontinent lithospheric weaknesses during the Sauk transgression and final Rodinia breakup.

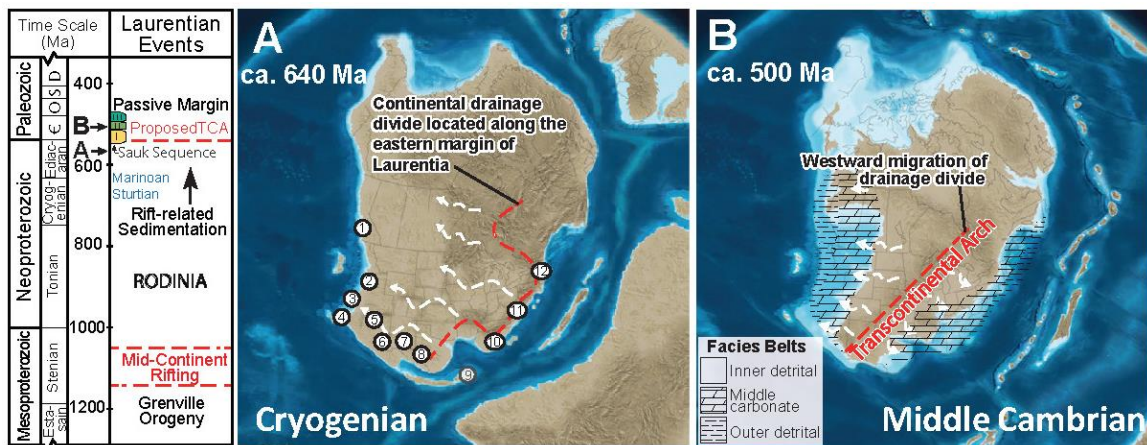
## 6.1. Introduction

Many detrital zircon datasets from extensional (rift, post-rift passive margin and intracratonic) basins show provenance shifts which are commonly interpreted to represent the effects of far-field tectonics that result in reorganization of continental drainage divides (Cawood et al., 2012; Gehrels & Pecha, 2014). Several detrital zircon provenance studies of western Laurentian Neoproterozoic and early Palaeozoic strata deposited during the Sauk transgressions (Sloss, 1963) propose that uplift of a northeast to southwest (all directions in modern coordinates) continental drainage divide across the midcontinent of Laurentia (the Transcontinental Arch; TCA; Fig. 6.1) resulted in the transition from Stenian (ca. 1.2–1.0 Ga) source terranes east of the TCA (Llano Uplift/Wichita province and/or Grenville) for Neoproterozoic strata, to more proximal, predominantly late Palaeoproterozoic source terranes west of the TCA for middle Cambrian through pre-Carboniferous strata (e.g. Amato & Mack, 2012; Gehrels & Pecha, 2014; Linde et al., 2014; Matthews et al., 2018; Mueller et al., 2007; Rainbird et al., 2017; Yonkee et al., 2014). The proposed timing of the initial uplift of the TCA varies from early Neoproterozoic (Craddock et al., 2017), late Neoproterozoic (Gehrels & Pecha, 2014), Cambrian (Linde et al., 2014; Saylor & Sundell, 2021), to Devonian (Myrow et al., 2003). Others (e.g. Howard et al., 2015; Karlstrom et al., 2018) have advocated that a combination of localized factors, predominately (Neoproterozoic–Cambrian) rift-related exhumation and reworking of underlying strata during the Sauk Transgression, require further consideration as potential mechanisms for this provenance shift. Here, we evaluate these competing models with a two-dimensional (U-Pb/Lu-Hf) quantitative comparison (Saylor & Sundell, 2021; Sundell & Saylor, 2021) of ca. 1.3–1.0 Ga detrital zircon components in Neoproterozoic to Cambrian strata of south-western Laurentia and Stenian-age sources to the east (e.g. Llano Uplift/Wichita province, Pikes Peak Batholith and/or Grenville province). We also revisit the timing of the provenance shift in question considering revision of the depositional ages of Cambrian strata in south-western Laurentia (Karlstrom et al., 2018, 2020; Spencer et al., 2014). Collectively, the results suggest a probable progressive shift in Stenian-age sources during the Cryogenian–Ediacaran, prior to a diachronous transition to predominantly late Palaeoproterozoic detrital zircon age components during the Cambrian. We suggest this provenance pattern

reflects progressive northeast to southwest uplift (or relative uplift) of a segmented TCA, which correlates with the geometry and timing of the Neoproterozoic–Cambrian rift-drift transition along the Iapetan margin.

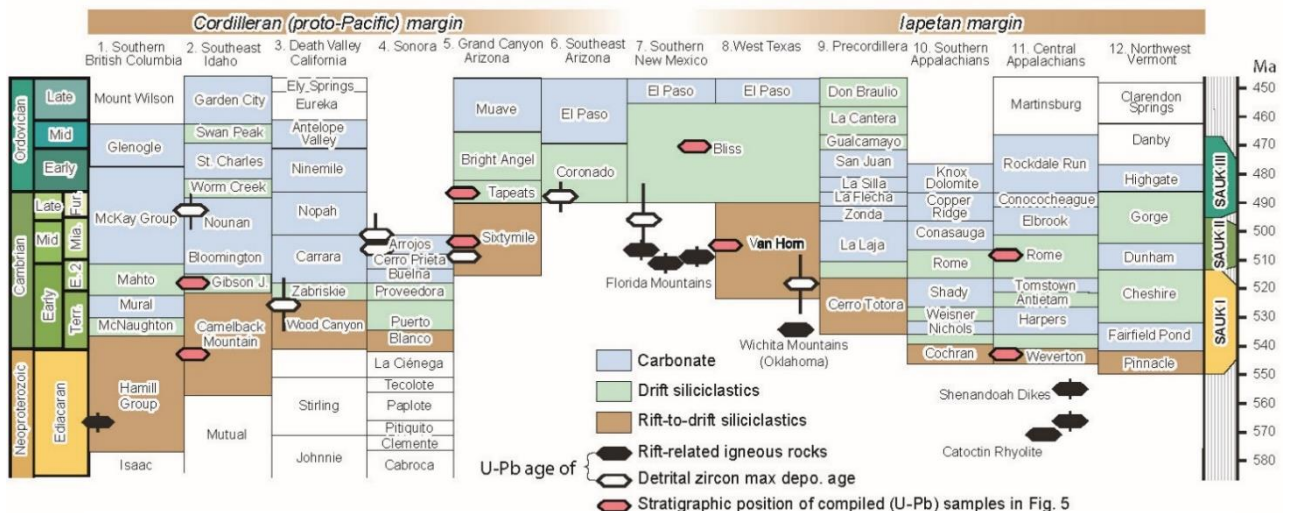
## 6.2. Laurentian Rifting and the Sauk Transgression

Sloss (1963) recognized that an extensive latest Precambrian to early Ordovician, unconformity-bounded stratigraphic “Sauk” transgressive mega-sequence occurred across most of Laurentia. The Neoproterozoic to early Ordovician Sauk mega-sequence has been further subdivided into three Sauk sub-sequences (or super-sequences; Keller et al., 2012; Palmer, 1981). These three sub-sequences (Sauk I, Sauk II and Sauk III) progressively onlap the cratonic platform and are separated by two regional unconformities, likely relating to relatively short-lived eustatic events at ca. 515 and 495 Ma respectively (Fig. 6.2; Burgess, 2019; Karlstrom et al., 2020; Keller et al., 2012; Palmer, 1981). However, a young and rapid ca. 505 to 501 Ma Sauk II transgression proposed for south-western Laurentia (Karlstrom et al., 2018) implies diachroneity both in the onset times of Sauk sequence transgressions and in drift-phase subsidence of the Laurentian rift margin, which may indicate important local tectonic influences in addition to global eustasy.



**Fig. 6.1:** Timeline for the Proterozoic and Early Palaeozoic major tectonic events for Laurentia with suggested palaeogeographic reconstructions at (a) ca. 640 Ma, showing the continental divide following the Grenville Belt along eastern Laurentia inferred from detrital zircon populations in Cryogenian–Ediacaran strata along the western margin and (b) at ca. 500 Ma, after relative uplift of the Transcontinental Arch and resulting westward migration of the continental divide resulting in a switch to more proximal source terranes for western Laurentia strata. Cambrian regional facies belts of the Sauk sequence are from Myrow et al., (2012). Numbered locations refer to locality of stratigraphic time–space diagrams in Fig. 6.2. The position of Sonora is restored along the Mojave mega-shear as per Stewart (2005) and Precordillera restored within the Ouachita embayment of per Martin, Collins, et al., (2020). Palaeogeographic reconstructions are modified from Gehrels and Pecha (2014) and Ron Blakey (<http://jan.ucc.nau.edu/rcb7/nam.html>), which are artistically rendered.

It is now recognized that the Sauk transgressive mega-sequence is likely related to the final breakup of Rodinia (Bond et al., 1984; Dalziel, 2014; Valentine & Moores, 1970). Palaeogeographic models of supercontinent Rodinia (e.g. Li et al., 2008) indicate a diachronous breakup, beginning along the western Laurentian margin with the emplacement of the ca. 780 Ma Gunbarrel magmatic event (Harlan et al., 2003). Related localized sedimentation occurred shortly after (Brennan et al., 2021; Dehler et al., 2017; Lund et al., 2003; Karlstrom et al., 2000) and along most of the western margin by ca. 720 Ma (e.g. Yonkee et al., 2014), coeval with the emplacement of the Franklin large igneous province (Macdonald et al., 2010). The western margin of Laurentia also records a younger (Ediacaran–Cambrian) phase of extension, breakup and syn-to post-breakup magmatism (Beranek, 2017; Bond et al., 1985; Prave, 1999; Ross, 1991).



**Fig. 6.2:** Generalized stratigraphic time–space diagram representative of the Neoproterozoic to Ordovician sedimentary units along the Laurentian margin adapted from Spencer et al., (2014). The ages of the reported maximum depositional ages (in Ma) are plotted as additional depositional constraints. Uncertainties are  $2\sigma$ . See Fig. 6.1 for generalized stratigraphic section locations. Red zircon symbols indicate the stratigraphic positions of compiled samples in Fig. 6.5. Various rock types, general depositional settings and indicated rift-drift intervals are per the following references: Southern British Columbia and Southeast Idaho (Colpron et al., 2002; Link et al., 1987; Yonkee et al., 2014); rift-related igneous rock age (Colpron et al., 2002); Worm Creek Quartzite detrital zircons (Link et al., 2017); Death Valley stratigraphy (Corsetti & Kaufman, 2003; Heaman & Grotzinger, 1992; Hogan et al., 2011; Prave, 1999); Death Valley detrital zircons (Stewart et al., 2001); Sonora stratigraphy (Farmer et al., 2005; Sour-Tovar et al., 2007; Stewart, 2005; Stewart et al., 2002); Grand Canyon stratigraphy (Karlstrom et al., 2000, 2018, and references therein); southeast Arizona stratigraphy (Hayes, 1972); south-central and southeast Arizona detrital zircons (Stewart et al., 2001); southern New Mexico stratigraphy (Hayes, 1972); southern New Mexico detrital/igneous zircons (Amato & Mack, 2012); west Texas stratigraphy (Hayes, 1972; LeMone, 1969); west Texas detrital zircons (Spencer et al., 2014); Precordillera stratigraphy (Finney et al., 2005); southern Appalachians stratigraphy (Chakraborty et al., 2012; Tull et al., 2010); central Appalachians stratigraphy (Astini, 1995; Burton & Southworth, 2010; Southworth et al., 2009; Tollo 2010); central Appalachians rift-related zircons (Aleinikoff et al., 1995, recalculated by Burton & Southworth, 2010; Southworth et al., 2009) and northwest Vermont stratigraphy (Brink et al., 2019; Landing et al., 2009; Shaw, 1958). Sauk sequence after Palmer (1981). Geologic time scale is after Peng et al., (2020) with Early, Middle and Late Cambrian divisions from the United State Geologic Survey time-scale (U.S. Geological Survey Geologic Names Committee, 2010).



Neoproterozoic to lower Palaeozoic strata preserved along the Laurentian margins were likely deposited within the proximal margin domain (e.g. Brennan et al., 2020; Yonkee et al., 2014). While precise timing is often difficult to determine, in proximal margin domains, the lithospheric breakup surface (i.e. rift-drift transition) is typically recorded as an unconformity (Soares et al., 2012). Evidence for a rift-drift transition at ca. 540 Ma in south-eastern Idaho (Linde et al., 2014; Yonkee et al., 2014) is recognized as an unconformity within the upper Brigham Group, where the feldspathic Ediacaran–Cambrian Camelback Mountain Quartzite disconformably overlies the quartzose Ediacaran Mutual Formation, delineating the base of the Sauk I super-sequence (Link et al., 1987). Further south, an interpreted rift-drift transition in the Grand Canyon region occurred at ca. 505 Ma, recorded as an unconformity between the coarse Cambrian Tapeats Sandstone and the underlying cherty siltstone and sandstone of the Ediacaran–Cambrian Sixtymile Formation. This unconformity is interpreted to coincide with the Sauk I/Sauk II boundary (Amato & Mack, 2012; Karlstrom et al., 2018; Spencer et al., 2014).

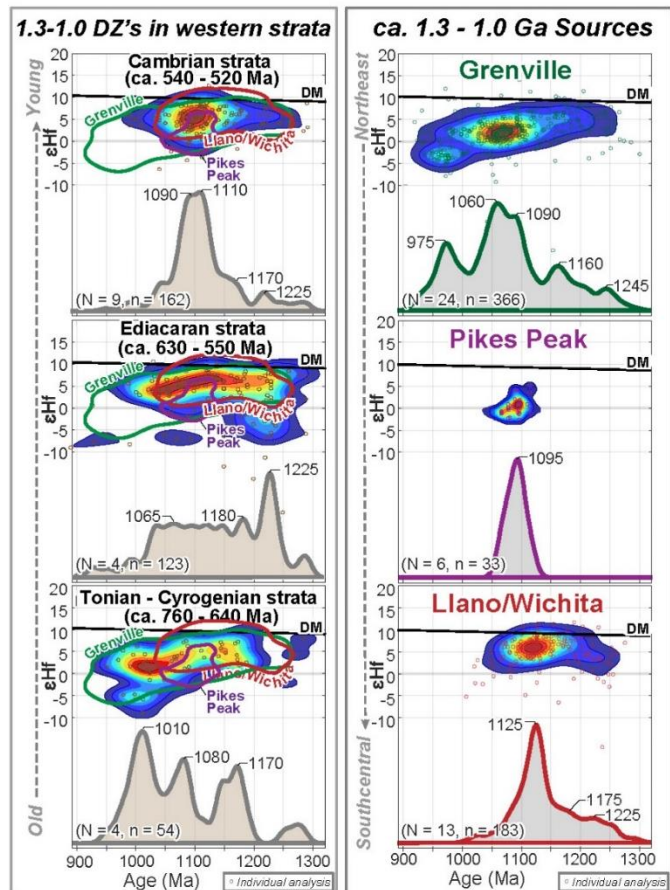
Provenance studies document an abundance of distal Stenian detrital zircons in Neoproterozoic strata and a provenance shift to more proximal, predominantly older sources in Cambrian strata (Linde et al., 2014; Matthews et al., 2018; Yonkee et al., 2014). The loss of Stenian detrital zircon populations in Cambrian strata has been interpreted as ca. 540–530 Ma chronostratigraphic indicator in Nevada, Utah, Idaho and Montana (Brennan et al., 2020; Gehrels & Pecha, 2014; Linde et al., 2014; Matthews et al., 2018; Yonkee et al., 2014), generally coinciding with initiation of the Sauk I transgression. However, the absence of biostratigraphic markers, and relatively limited radiometric age control due to the scarcity of syn-depositional volcanism, implies that the depositional ages of these clastic rocks may be poorly constrained and so the inferred synchronicity of the provenance changes remains an open question.

### **6.2.1 Evaluating the Stenian sources**

Until recently, quantitative assessment of detrital zircon distributions and potential source similarities/dissimilarities was limited to one-dimension, often (U-Pb) age-only, statistical comparison methods (e.g. Nordsvan et al., 2020; Saylor & Sundell, 2016; Vermeesch, 2012, 2013, 2018). However, it is difficult to differentiate between separate source regions with similar magmatic histories using these methods. Consequently, we compiled bivariate (U-Pb and Lu-Hf) datasets (Gehrels & Pecha,

2014; Howard et al., 2015; Martin, Spencer, et al., 2020; Spencer et al., 2012; Wooden et al., 2013 and references therein) of ca. 1,300 to 900 Ma detrital zircon components in late Tonian (<760 Ma) to Cambrian strata of western/south-western Laurentia. These strata are compared to potential source regions in eastern Laurentia (Grenville province), south-eastern Laurentian (Llano and Wichita province) and central Laurentia (Pikes Peak Batholith; Fig. 6.3) utilizing recent two-dimensional quantitative comparison methods (DZstats2D MATLAB code; Sundell & Saylor, 2021). Cross-correlation coefficient results (Fig. 6.4) indicate that north-eastern Stenian-age sources (Grenville) show the strongest statistical correlation with the detrital zircon components in the oldest Tonian–Cryogenian strata, and generally show decreasing similarity with younger Ediacaran and Cambrian strata. Conversely, the similarity of south-western Laurentian strata with south-eastern Stenian-age sources (Llano and Wichita province) increases up-section and is strongest in Cambrian strata. Both the Ediacaran and Cambrian strata show a second strongest similarity with their immediately underlying (older) Tonian–Cryogenian or Ediacaran strata respectively. Additional comparison metrics (Similarity, Likeness, K-S Test D and Kuiper Test V) all show the same trend as the Cross-correlation metric (see supplementary information).

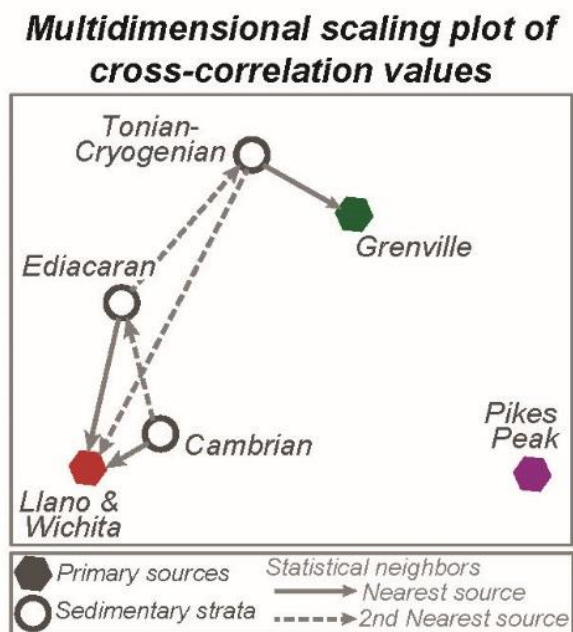
**Fig. 6.3:** Approximately 1.3–0.9 Ga detrital zircon U-Pb/ Lu-Hf values from (ca. 760–640 Ma) Tonian–Cryogenian, (ca. 630–550 Ma), Ediacaran (ca. 540–530 Ma) and Cambrian strata in south-western Laurentia and potential ca. 1.3–0.9 Ga eastern source provinces including the Grenville (eastern Laurentia), Pikes Peak (central Laurentia) and Llano/Wichita province (south-eastern Laurentia; see Fig. 6.6). Age (U-Pb) data are shown in kernel density estimation plots with 10 m.y. bandwidths generated with DensityPlotter (Vermeesch, 2012). Lu-Hf data are shown in with individual analysis scatter plots, and bivariate kernel density estimation plots with 25 million-year bin and 1.5 εHf bandwidths generated with the MATLAB code of Spencer et al., (2020). Abbreviations: DM, depleted mantle; N, number of samples; n, number of analyses.



## 6.2.2 A diachronous loss of Stenian zircon grains?

In south-eastern Idaho and northern Utah (location 2 in Figs. 6.1, 6.2, 6.5 and 6.6), the provenance shift from Stenian sources to predominantly older sources occurs within the Camelback Mountain Quartzite that spans the Ediacaran–Cambrian boundary (Link et al., 1987; Yonkee et al., 2014). Approximately 2,500 meters stratigraphically above where this provenance shift occurs, feldspathic sandstones deposited within a predominately carbonate sequence mark the Sauk II/Sauk III sequence boundary (Armstrong & Oriel, 1965). These feldspathic sandstones (Worm Creek Quartzite Member of the St. Charles Formation) have a late Cambrian (Furongian) age of ca. 495 Ma (detrital zircon maximum depositional age of ca. 497 Ma and Dunderbergia and Elvinia trilobite zones from 495.2 to 493 Ma; Link et al., 2017).

**Fig. 6.4:** Multi-dimensional scaling (MDS) plot of two-dimensional (U-Pb and Lu-Hf) cross-correlation comparison results for ca. 1.3 to 0.9 Ga zircon components. MDS plots are used to visualize the similarity of datasets, where similar datasets plot closer together in Cartesian space, and less similar datasets plot further apart (see Nordsvan et al., 2020; Saylor & Sundell, 2021; Vermeesch, 2013, for further discussion of MDS plots as applied to zircon datasets). Statistically nearest (or most similar) sources (solid arrow) and second nearest (dashed arrow) geologically feasible sources are indicated. Geologically feasible indicating that it is available as a source at the time of the units' deposition (e.g. Ediacaran strata cannot be sourced from recycling of younger Cambrian strata). Note that Tonian–Cryogenian strata show the greatest similarity to a Grenville source, while Ediacaran and Cambrian strata show a greatest similarity to a Llano and Wichita source, with a second greatest similarity to Tonian–Cryogenian and Ediacaran strata respectively. Similarity, Likeness, Kolmogorov–Smirnov Test D and Kuiper Text V comparison metrics were also calculated and all show the same statistically nearest and second nearest sources indicating reliability of the results. Comparison matrices for all metrics were generated with the DZstats2D MATLAB script of Saylor and Sundell (2021).



In Arizona and west of the TCA (location 5 in Figs. 6.1, 6.2, 6.5 and 6.6) within similar-age strata (Spencer et al., 2014), deposition of the ca. 540–512 Ma Sixtymile Formation spans the Sauk I transgression and contains significant Stenian detrital zircon age components throughout its various stratigraphic intervals (Karlstrom et al., 2018). The provenance shift from dominantly Stenian detrital zircon sources to proximal, dominantly older, sources is not observed in this sequence until the younger (<508 Ma) Tapeats Sandstone, which was deposited during the second major Sauk II transgression (Karlstrom et al., 2018, 2020). The recognition of Cambrian-age detrital zircon in these rocks allowed precise maximum depositional age constraints via tandem dating: laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) of a large number of grains, followed by chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-TIMS) of the youngest grains (Karlstrom et al., 2018, 2020).

East of the TCA in south-eastern New Mexico and western Texas (location 7 in Figs. 6.1, 6.2, 6.5 and 6.6), rocks coeval with the Sixtymile Formation and Tapeats Sandstone that are also interpreted to span the Sauk I/Sauk II boundary contain significant Stenian detrital zircon populations (Amato & Mack, 2012; Spencer et al., 2014). Also east of the TCA, in the central Appalachian region (location 11 in Figs. 6.1, 6.2, 6.5 and 6.6), the Weverton Formation and the overlying Rome Sandstone also contain significant Stenian detrital zircon components. These rocks are interpreted to record the basal Sauk I transgression, suggesting they are mostly similar in age to the upper Brigham Group of southeast Idaho, i.e. Camelback Mountain Quartzite and Gibson Jack Formation (Fig. 6.2). The lower Cambrian Weverton Formation is generally interpreted to record provenance from proximal Grenvillian basement. The overlying Rome Sandstone contains Stenian detrital zircon in addition to older ages typical of the Laurentian craton (including the Granite-Rhyolite, Trans-Hudson, Penokean and Superior provinces), consistent with an Early–Middle Cambrian drainage system that originated in central Laurentia and flowed eastward (Thomas et al., 2004). Similar detrital zircon patterns are found further north in strata of comparable age along the rifted margin of Laurentia in Newfoundland (Cawood & Nemchin, 2001).

Although prolonged rifting started at ca. 750 Ma, continental break-up along the eastern Laurentian margin occurred later than along the western margin (Li et al., 2008; Merdith et al., 2021). Rift-to-drift magmatism indicates continental separation

along a north-to-south propagating rift, or distributed rift system, from ca. 615 to 510 Ma, resulting in the opening of the Iapetus Ocean and/or removal of further blocks into an already open Iapetus Ocean (Bond et al., 1984; Cawood et al., 2001; Dalziel, 2014; Mitchell et al., 2011). The rift-drift transition initiated in the central Appalachians at ca. 545 Ma and propagated southward to West Texas by ca. 510 Ma (Fig. 2.2; Martin et al., 2020; Spencer et al., 2014). Iapetan-associated rifting propagated far into the craton, forming intra-continental rift basins (or aulacogens), including the Reelfoot Rift and the Southern Oklahoma aulacogen (Fig. 6.6; Thomas, 2011).

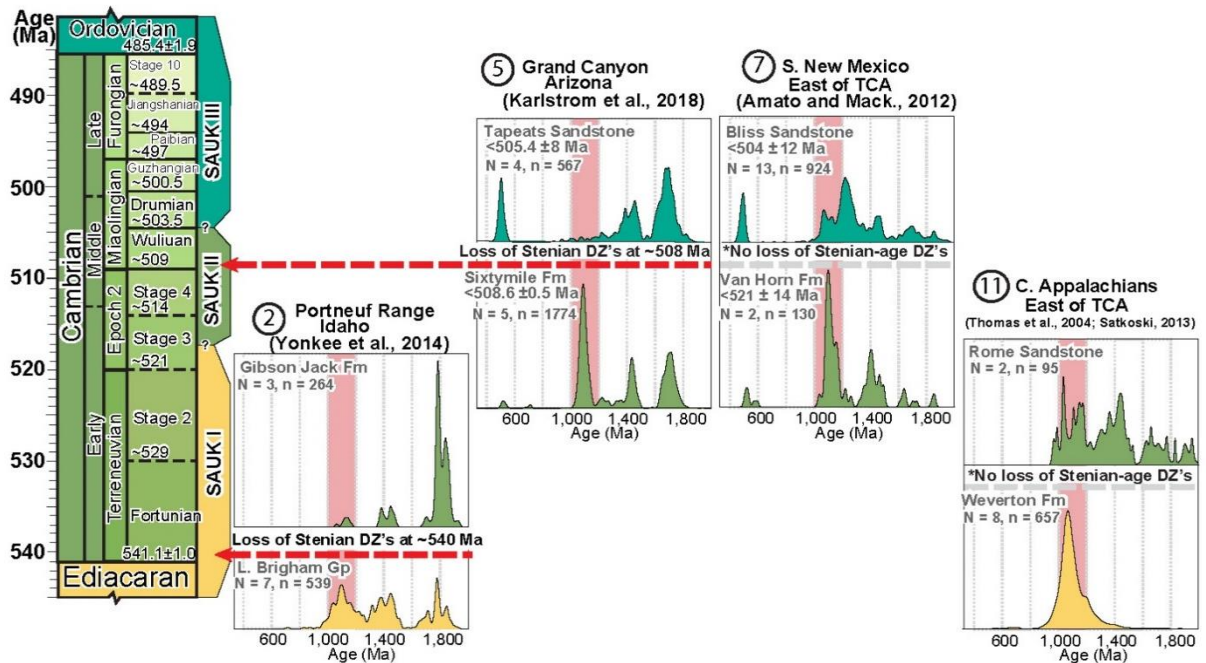
### **6.2.3 The Transcontinental Arch**

Initial isopach maps suggested that the Transcontinental Arch (TCA) was a southwest-trending elevated linear feature during Cambrian time (Sloss, 1988). However, stratigraphic compilations and further isopach mapping (Carlson, 1999; Poole et al., 1992) indicate that the TCA was likely a composite, broad and discontinuous plateau dissected by several smaller northwest-striking, laterally persistent, basement-controlled structures of the Mazatzal-Yavapai province highs that served as local sources of detritus for small intervening basins, resulting in a broad, yet segmented TCA (Carlson, 1999).

## **6.3. Tectonic Model: Progressive uplift of an arch between two young passive margins**

Bond et al., (1989) advocate that flexural bending of the Laurentian margin and long-term eustasy is responsible for the retreat of the siliciclastic shoreline during the Sauk transgression and have calculated broad flexural uplifts of 100–200 m located over 600 km inboard from the Cordilleran continental margin. These results suggest that sediment loading, thermal densification and post-rifting lithospheric rigidity increase along the craton margin could result in broad uplift of the midcontinent region. Widespread denudation occurred during the Neoproterozoic associated with the formation of the “Great Unconformity” (Flowers et al., 2020; Keller et al., 2018). This denudation, along with the lateral continuity in sedimentary thickness and facies of the Sauk sequence (Sloss, 1963), suggests that the sub-Sauk depositional surface across Laurentia exhibited rather low relief. For example, the Cambrian basal transgressive unit in Wyoming and Montana (the Flathead Quartzite) shows a remarkably uniform average thickness of approximately 35 meters over a region in

excess of 90,000 square kilometres (Deiss, 1935). Thus, we propose that a regionally extensive uplift (or relative uplift) of any significant magnitude would have had a notable influence on sedimentary transport systems.

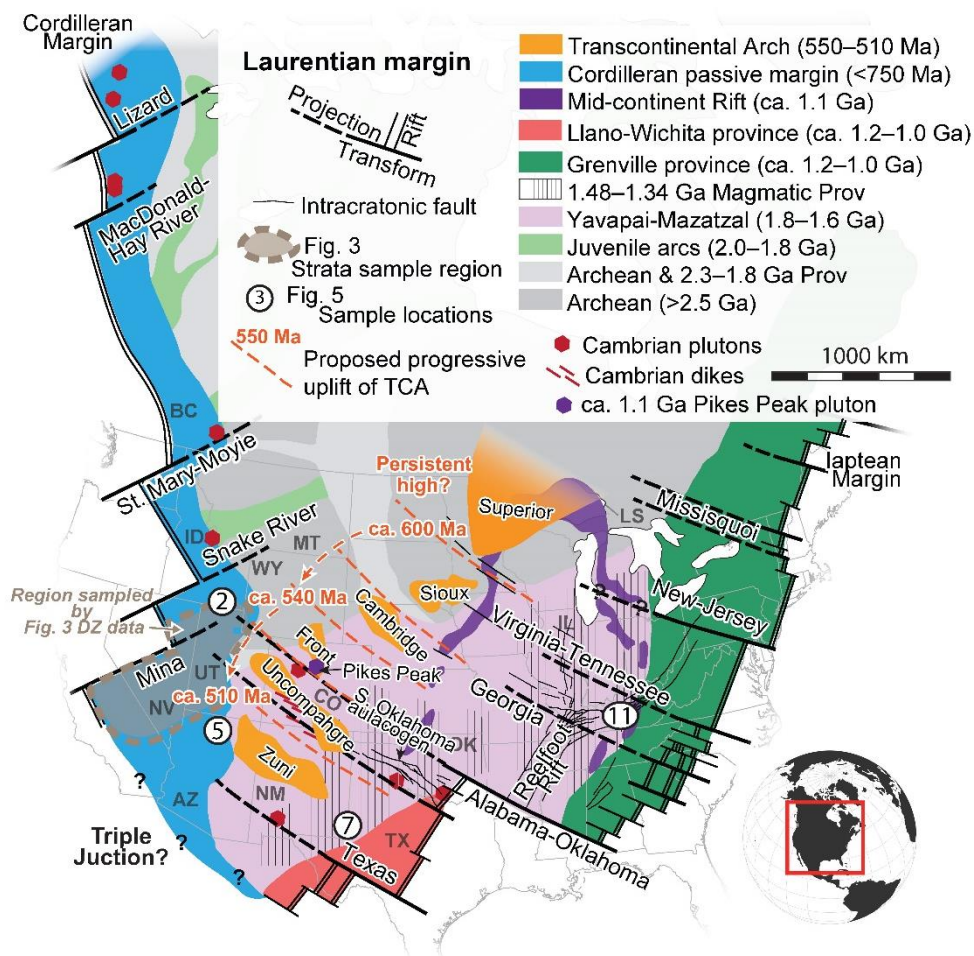


**Fig. 6.5:** Proposed timeline for the widespread provenance shift from dominant Stenian (ca. 1.2–1.0 Ga) detrital grains to older dominantly Palaeoproterozoic detrital grains. In Idaho, northern Utah and Nevada (location 2), this shift has been previously stratigraphically recognized to take place during the Sauk I transgression at ~540–530 Ma (Linde et al., 2014; Yonkee et al., 2014). To the south in the Grand Canyon region (location 5), this shift is precisely placed (by  $508.6 \pm 0.8$  TIMS max depositional age on youngest detrital zircon in Sixtymile Formation) to the Sauk II transgression at ~508 Ma (Karlstrom et al., 2018). Red dashed lines indicate the approximate timing of the provenance shift. East of the Transcontinental Arch at location 7 (Amato & Mack, 2012; Spencer et al., 2014), in a stratigraphic section similar in age to the Grand Canyon section, no loss of Grenville detritus is observed. Also east of the Transcontinental Arch at location 11 (Thomas et al., 2004; Satkoski, 2013), in a stratigraphic section generally similar in age to the Idaho section, no loss of Stenian detrital zircons and possibly a shift to Laurentian midcontinent sources is observed. Data are shown in normalized kernel density estimation plots with 10 m.y. bandwidths. Full sample list can be found in supplementary material A. DZ, detrital zircon;  $N$ , number of samples;  $n$ , number of analyses.

At first order, comparison of bivariate (U-Pb and Lu-Hf) datasets suggests that Neoproterozoic–Cambrian strata in south-western Laurentia record a decrease in the contribution of Stenian zircon from north-eastern sources (Grenville) with time and a corresponding increase in the contribution of south-eastern Stenian sources (Llano and Wichita provinces), with recycling from proximal underlying strata being a secondary contributor. Furthermore, the stratigraphic interval that contains the switch from predominantly Stenian zircon components to older Palaeoproterozoic components is variable. In south-eastern Idaho/northern Utah, the shift occurred during the Sauk I



transgression at ca. 540 Ma, whereas in south-eastern New Mexico/western Texas Stenian zircon components are present throughout Sauk I strata, and the shift does not occur until ca. 510 Ma within Sauk II transgressive strata. These trends appear to be consistent with progressive (northeast to southwest) west-stepping migration of a significant continental drainage divide resulting from Ediacaran to Cambrian uplift of the TCA. This drainage divide migration probably cut off westward transport of distal Grenville sourced zircon to western/south-western Laurentia at ca. 540 Ma, while northern/north-western transport of more proximal Llano and Wichita sourced zircon appears to have persisted until ca. 510 Ma. This provenance pattern likely reflects progressive (northeast to southwest) formation of the TCA during the Ediacaran–Cambrian resulting from the subsidence of passive margins (and perhaps intra-continental rift basins such as the Reelfoot Rift) surrounding Laurentia, and the resulting lithospheric flexure during the breakup of Rodinia.



(Previous page) **Fig. 6.6:** Map of the tectonic structures suggested to be interrelated and have influenced the uplift of the Transcontinental Arch (TCA). Laurentian basement provinces and structures adapted from Whitmeyer and Karlstrom (2007), Gehrels and Pecha (2014), Marshak et al., (2017) and Stein et al., (2018). The ca. 1.1 Ga Midcontinent Rift (Stein et al., 2018), the Iapetus rift margin (Thomas, 2014), the western Laurentian margin (Lund, 2008) and uplifted regions of the TCA (Carlson, 1999) all show geometric affinities with each other. Cambrian plutons/dikes/localized detrital zircon age components from Lund (2008); Amato and Mack (2012); Hanson et al., (2013); Link et al., (2017) and references within. Iapetan transform faults bisect segments of the Midcontinent rift and the TCA. Although precise timing of the rift-drift transition along the Iapetan margin is difficult to determine, generally it is thought to have progressively opened from north to south at ca. 615 to 500 Ma (Dalziel, 2014); with rift-drift transitions in the central Appalachians at ca. 545 Ma, and west Texas at ca. 510 Ma (Spencer et al., 2014), consistent with our proposed timing of northwest to southeast progressive uplift of the Transcontinental Arch. The proposed geometry of the Laurentian margin is consistent with an oceanic triple-junction located off south-western Laurentia. AZ, Arizona; BC, British Columbia; IL, Illinois; OK, Oklahoma; NV, Nevada; NM, New Mexico; MT, Montana; ID, Idaho; UT, Utah; WY, Wyoming; TX, Texas; LS, Lake Superior.

The major transverse faults and fault zones of the Iapetan margin (New Jersey, Virginia-Tennessee, Georgia, Alabama-Oklahoma and Texas; Thomas, 2014) align with segments of the Midcontinent Rift and the proposed north-eastward-trending tectonic units of the TCA (Fig. 6.6). Localized Cambrian deformation, dike swarms and associated magmatism are identified more than 800–1400 km inwards from the continental margin, adjacent to these major transverse fault zones in Colorado, northern New Mexico and perhaps central Illinois (Fig. 6.6; Brueseke et al., 2016; Freiburg et al., 2020; Hanson et al., 2013; Thomas, 2014). These observations lead us to link these Laurentian structures and suggest that the segmented geometry of the Mesoproterozoic Midcontinent Rift acted as a genetic control of the segmentation of the Neoproterozoic Iapetan rifted margin and influenced the geometry of alternating broad highs and intervening lows of the TCA. These transverse fault zone/transverse fault zone-parallel structures initially formed during Midcontinent rifting ca. 1.1 Ga, and we suggest were rejuvenated during Neoproterozoic–Cambrian opening of the Iapetus Ocean and (relative) uplift of the Transcontinental Arch, prior to probable reactivation during Ordovician closure of the Iapetus and subsequent late Palaeozoic closure of the Rheic Ocean (e.g. Craddock et al., 2017). Our analysis suggests that intra-continental transverse rift-structures are long-lived regions of lithospheric weakness.



## 6.4. References

*Terra Nova reference style*

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# Chapter 7: Synthesis and Conclusions

## 7.1 Introduction

The geologic focus of this thesis spans over 2,000 million years, from the Neoproterozoic Era to the Cambrian Period. During this time span, Earth underwent numerous cyclical, secular and stepwise modifications. To espouse the preamble quote of this thesis, there are many “horizons” during these 2,000 million years where the geological evidence, and consequently our knowledge of these Earth modifications is missing or incomplete. This work aimed to reduce the extent and uncertainty of several key tectonic/geologic horizons within the key piercing point of the Belt Basin region of western Laurentia, and thus increase the completeness of the Paleoproterozoic to early Paleozoic knowledge of this critical area in many tectonic reconstructions. In doing so, this work provides a more robust regional geologic history against which inherently more speculative larger scale global Nuna and Rodinia tectonic models must be compared, evaluated and redefined.

In the following chapter, the work of this thesis is synthesized and discussed within the context of three general time intervals that correspond to significant global plate-tectonic reorganization events. These time intervals include: ca. 2000–1350 Ma (section 7.2.1) during which supercontinent Nuna assembled, ca. 1350–900 Ma (section 7.2.2) during which supercontinent Nuna broke apart and subsequent supercontinent Rodinia assembled, and finally from ca. 800–500 Ma (section 7.2.3) during which Rodinia broke apart. In a way, these three delineations are a Belt Basin regionally focused update to the unconformity bounded successions A (~1700–1200 Ma), B (1200–800 Ma), and C (800–570 Ma) defined by Young et al., (1979) in northern Laurentia, and subsequently applied southward to most of the western Laurentian margin (Link et al., 1993).

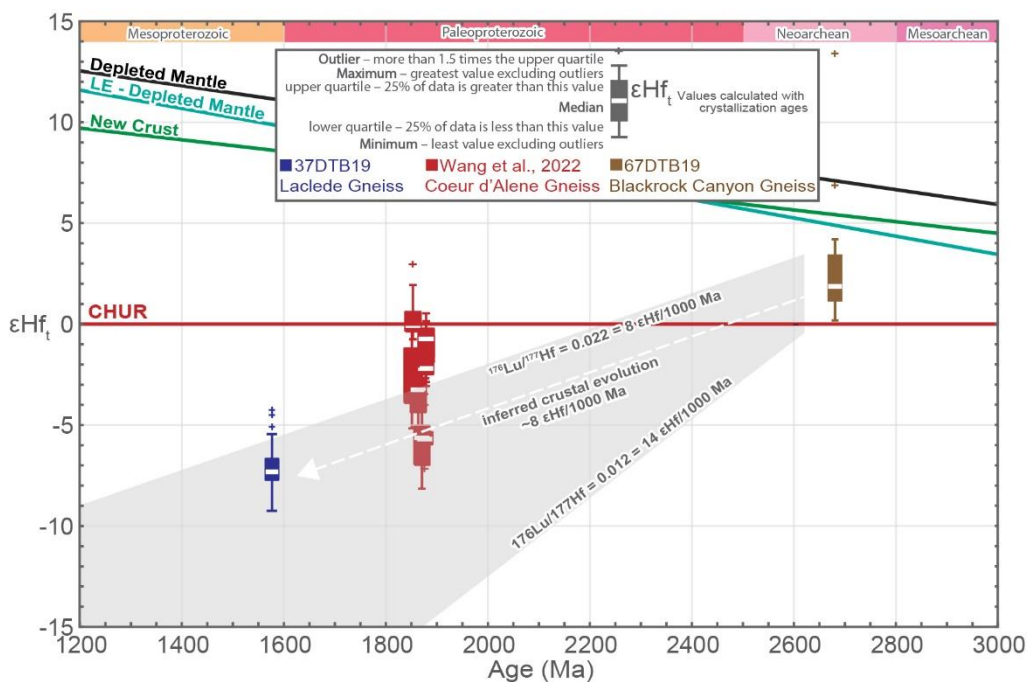
## 7.2 Summary and significant findings

### 7.2.1 Paleoproterozoic Laurentia and Nuna assembly

As introduced in Chapter 1, a main controversy surrounding global Paleoproterozoic paleogeography is whether the final assembly of supercontinent Nuna was primarily a ca. 1800 Ma process coeval with the main assembly of Laurentia (e.g. Zhao et al., 2002), or if final Nuna assembly was primarily a later, ca. 1600 Ma process (e.g. Kirscher et al., 2021). In northwestern Laurentia and northeastern Australia, ca. 1750–1550 Ma strata record final Nuna assembly at ca. 1600 Ma

(Furlanetto et al., 2016; Nordsvan et al., 2018). However, ~1500 km further south in the Belt Basin region, the focus area of this study, previous provenance studies have primarily focused on the ca. 1470–1380 Ma Belt Supergroup (e.g. Ross and Villenaueve, 2003; Lewis et al., 2007) and neglected the significance of potential pre-Belt Supergroup units, such as the Gold Cup and likely correlative Neihart quartzites (Doughty and Chamberlain, 2008) to constrain the tectonostratigraphic record of this region prior to ca. 1470 Ma.

As discussed in Chapter 3 (section 3.2.1) the Mesoproterozoic Belt Basin traverses at least two significant Paleoproterozoic sutures associated with the assembly of Laurentia, which include the Great Falls Tectonic Zone and the Vulcan Structure. The better exposed, and thus better studied of these structures is the Great Falls Tectonic Zone which records ca. 1860–1730 Ma convergent events associated with northwest dipping subduction and amalgamation of the Medicine Hat and likely Clearwater block(s) to the Wyoming province (e.g. Mueller et al., 2002; Foster et al., 2006; Gifford et al., 2020). In Chapter 2, existing (Vervoort et al., 2016) and new zircon geochronology from the Clearwater Block indicates that crustal growth in this region occurred primarily at ca. 2670 and ca. 1880–1840 Ma, with localized magmatism at ca. 1581 Ma. New zircon Lu/Hf and O- isotope data, also suggest that the younger interval of Paleoproterozoic crustal growth primarily records remelting of older lithosphere, as opposed to significant juvenile mantle contributions. However, additional constraints on this period of ca. 1880–1840 Ma crustal growth in the Clearwater block have been published (Wang et al., 2022) following the publication of Chapters 2 and 3. Thus, we include these new results here (Fig. 7.1).



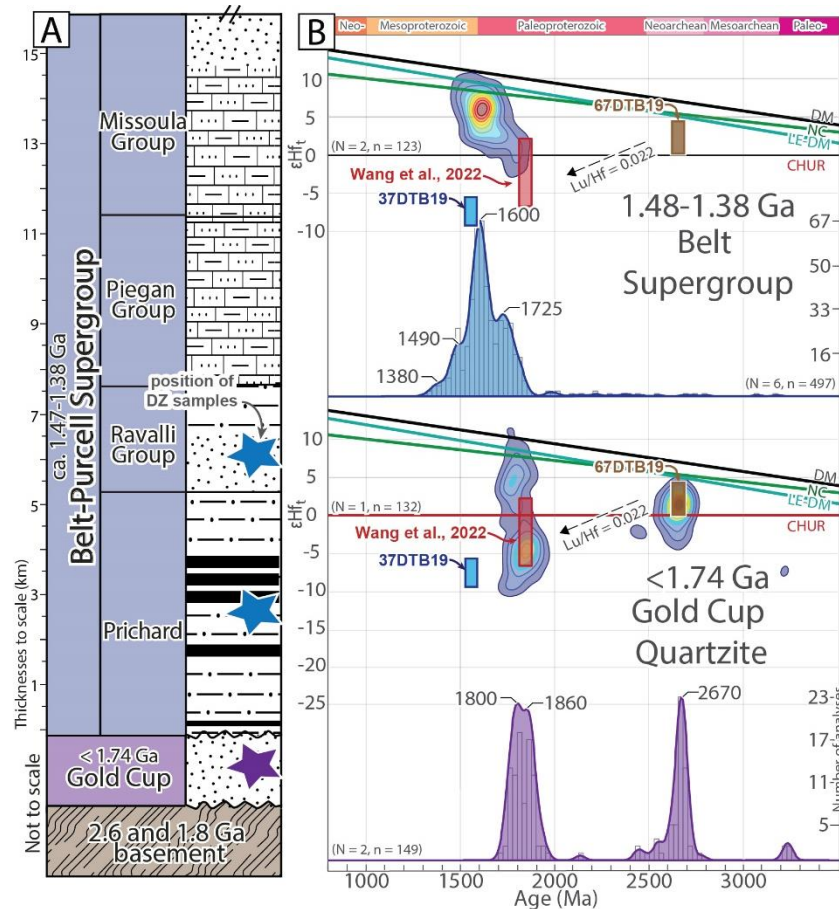
**Fig. 7.1:**  $\epsilon\text{Hf}_t$  values vs.  $(^{207}\text{Pb}/^{206}\text{Pb})$  Age plot for the Clearwater Block (Chapter 2; Wang et al., 2022). The colored box plots indicate the range of  $\epsilon\text{Hf}_t$  values calculated for each gneiss based on the measured Lu/Hf, and the interpreted crystallization age of the sample. See Fig. 2.7 for explanation of Mantle evolution curves. The red boxes are from the Coeur d'Alene Gneiss in the Priest River complex near the Laclede Gneiss outcrop (Wang et al., 2022). Wang et al., (2022) also reports Lu/Hf data from ca. 1900–1830 Ma orthogneiss located further south in the Clearwater complex, which show a slightly wider range of  $\epsilon\text{Hf}_t$  values from approximately  $-8$  to  $+8$  (Wang et al., 2022). CHUR—Chondritic Uniform Reservoir (Bouvier et al., 2008).

In Chapter 3, additional U/Pb (prior limited n-analyses presented by Ross and Villeneuve, 2003; Doughty and Chamberlain, 2008; Mueller et al., 2016) and the first Lu/Hf detrital zircon results from the oldest, younger than ca. 1750 Ma strata that unconformably overlie the Clearwater and Medicine Hat/Great Falls Tectonic Zone basement terranes, the Gold Cup Quartzite and Neihart Quartzite, was presented. The objective of this data was to evaluate whether these likely pre-Belt Supergroup units require non-Laurentia sources or not, and thus assess competing models on the timing of Nuna assembly. The detrital zircon components of both the Gold Cup and Neihart quartzites consist of primarily two age distributions at ca. 2670 Ma and ca. 1860–1780 Ma. The Neoproterozoic grains contain CHUR-like Lu/Hf values, and consequently are similar in age and Hf composition to the nearby Clearwater Block basement. However, the Paleoproterozoic grains contain a wider range of Lu/Hf values that are generally concentrated around three  $\epsilon\text{Hf}$  ranges,  $+8$  to  $+4$ ,  $-3$  to  $-7$ , and  $-14$  to  $-18$  (Neihart Quartzite only), suggesting perhaps mixing of at least three possible sources. Suitable sources for these grains are found within ca. 1800 Ma ( $\epsilon\text{Hf}_t = -10$  to  $-20$ ) rocks of the

Medicine Hat block (Gifford et al., 2020), ca. 1880–1840 Ma ( $\epsilon\text{Hf}_t = +8$  to  $-8$ ) rocks of the Clearwater complex (Vervoort et al., 2016; Wang et al., 2022), the extensive ca. 1860–1780 Ma Trans-Hudson orogen, and perhaps ca. 1880–1740 Ma ( $\epsilon\text{Hf}_t = +10$  to  $-5$ ) rocks of the Mojave-Yavapai-Mazatzal provinces (Whitmeyer and Karlstrom, 2007; Holland et al., 2018 and references within). Consequently, the tectonostratigraphic record of the Belt region does not require any non-Laurentian sources during Laurentian assembly, and thus is broadly consistent with the ca. 1800 Ma assembly of the main Archean blocks (Wyoming, Medicine Hat, Clearwater, Hearne) of western Laurentia prior to a later ca. 1650–1600 Ma assembly of Nuna. In western/southwestern Laurentia, a record of this later assembly may be associated with ca. 1670 Ma metamorphism in the Farmington Zone (Mueller et al., 2011), and the ca. 1650–1600 Ma Mazatzal Orogeny (Gibson and Champion, 2019; Holland et al., 2020).

In Chapter 2, the  $1581 \pm 3$  Ma Laclede Gneiss within the Priest River complex contains evolved Lu–Hf ( $\epsilon\text{Hf} = -6$  to  $-10$ ) and mantle-like  $\delta^{18}\text{O}$  zircon values, suggesting remelting of a Neoproterozoic lower continental lithosphere source. The evolved values of this gneiss makes it a poor direct match for coeval, mostly juvenile, magmatism in the Gawler Craton of southeastern Australia (Reid and Payne, 2017). Thus, I advocated that the Laclede Gneiss is not a stranded fragment of Australian crust. However, as Wang et al., (2022) note, Antarctica and southeastern Australia do have similar age (ca. 1580 Ma) evolved (negative  $\epsilon\text{Hf}_t$ ) magmatism that could be a non-Laurentian source for the Laclede Gneiss. Nevertheless, the evolved (negative  $\epsilon\text{Hf}_t$ ) values for the Laclede Gneiss importantly demonstrate that a non-Laurentian source for the similar-age evolved (positive  $\epsilon\text{Hf}$ ) detrital zircon grains within the surrounding Belt Supergroup is required. This zircon source is likely the Gawler Craton of southeastern Australia, as it contains magmatic rocks with the same age and Hf values as the juvenile ca. 1600 Ma detrital zircon grains within the overlying Belt Supergroup (Fig. 7.2). Ca. 1630–1580 Ma St. Peters Suite and Gawler Range Volcanics in southeastern Australia have plume and/or back-arc genesis models (e.g. Reid and Payne 2017; Chapman et al., 2019). Consequently, as discussed in Chapter 2 (section 2.5.4), the Laclede Gneiss may be a crustal-derived, distal manifestation of a ca. 1580 Ma mantle up-welling (perhaps plume) event beneath southeastern Australia. Therefore, the Belt Basin regional geologic record does not require the assembly of Australia/East Antarctica with western Laurentia within Nuna at ca. 1800

Ma, and rather the first indirect evidence of this association may be recorded in the 1581 Ma Laclede Gneiss.



**Fig. 7.2:** Paleoproterozoic stratigraphy and U–Pb/Lu–Hf detrital zircon results from the Priest River Region (Doughty and Chamberlain, 2008; Box et al., 2020; Lewis et al., 2020; Brennan et al., 2021) and the range of robust  $\epsilon\text{Hf}_t$  values (samples 67DTB19 and 37DTB19) from the underlying ca. 2670 and 1580 Ma igneous basement. The red box indicates the age and  $\epsilon\text{Hf}_t$  range (ca. 1.88–1.83 Ga) of Paleoproterozoic magmatism in the northern extent of the Clearwater Block (Vervoort et al., 2016) and the Lu–Hf values of Wang et al., (2022) for the Coeur d’Alene Gneiss. Wang et al., (2022) also reports Lu/Hf data from ca. 1900–1830 Ma orthogneiss located further south in the Clearwater complex, which show a slightly wider range of  $\epsilon\text{Hf}_t$  values from approximately –8 to +8, which could be the source of the more juvenile ca. 1860–1830 Ma grains in the Gold Cup Quartzite. (Wang et al., 2022). Inferred crustal evolution line (Lu/Hf = 0.022) for the Clearwater Block indicated. LE–Late Extraction, NC–New Crust, CHUR–Chondritic Uniform Reservoir. See Fig. 2.7 for explanation of mantle evolution lines.

### 7.2.2 Mesoproterozoic Laurentia and the Nuna to Rodinia transition

In Chapters 2 and 4, it was demonstrated that the ca. 1600 Ma detrital zircon grains within Belt Basin strata were not sourced from the Laurentian ca. 1581 Ma Laclede Gneiss, as its evolved (negative  $\epsilon\text{Hf}_t$ ) values could not have supplied the juvenile (positive  $\epsilon\text{Hf}_t$ ) detrital grains. Other workers (e.g. Yao et al., 2017; Xu et al., 2019; Goodge et al., 2017) have suggested that ca. 1570–1510 Ma magmatic rocks in Hainan Island (south China) or East Antarctica could have contributed detritus to the

Belt Basin. However, zircon grains from these magmatic rocks also exhibit evolved (negative  $\epsilon_{\text{Hf}_i}$ ) values (see Fig. 2.10), and thus are also poor matches for the grains in Belt Basin strata. Consequently, the mostly juvenile ca. 1630–1580 Ma magmatism within the Gawler craton of southeastern Australia is the best match for these detrital grains, adding further support to long-standing correlations between these two regions during the Mesoproterozoic (Ross et., 1992; Ross and Villeneuve, 2003).

A long-standing controversy in Belt Supergroup geology, is whether the ca. 1470–1380 Ma Belt Basin is a rift to passive margin sequence that records Nuna breakup at ca. 1400 Ma, or whether the Belt Basin was completely intracratonic and preceded Nuna breakup (e.g. Evans et al., 2000; Ross and Villeneuve, 2003; Lonn et al., 2020 and references within). In fact, paleomagnetic data indicates that Australia and western Laurentia were likely in a stable configuration from ca. 1600–1300 Ma suggesting that the Belt Basin may have been completely intracratonic and preceded ca. 1300–1200 Ma Nuna breakup (Kirscher et al., 2021). Limited ( $n = 29$  within their applied -5 to 20% discordant filter) U/Pb detrital zircon results from upper Belt Supergroup strata in northeastern Washington reported by Box et al., (2020) also suggests that the exotic, likely Australian, ca. 1630–1580 Ma source terrane shared sediments with Laurentia until ca. 1390 Ma. The limited geochronology results (of Box et al., 2020) from upper Belt Supergroup, Missoula Group strata (Snowslip Formation) in the Priest River region require substantiating, particularly due to the notable Pb-loss in this sample. It is worth noting that I analyzed a similar upper Belt Supergroup, Missoula Group (Snowslip Formation) sample (14DTB19; see Chapter 3), which yielded almost entirely discordant ages consistent with ca. 95 Ma Pb-loss. Nevertheless, the Missoula Group results of Box et al., (2020) do suggest an Australian-Laurentian connection within Nuna until at least ca. 1390 Ma. However, until recently (Chapter 3; Box et al., 2020), no strata between the ca. 1380 Ma upper Belt Supergroup and ca. 720 Ma basal Windermere Supergroup, were identified in the Belt Basin region. Consequently, there was a limited regional geologic record to compare against paleomagnetically constrained models of ca. 1300–1200 Ma Nuna breakup.

Early work on the Deer Trail Group in northeastern Washington State (Miller and Whipple, 1989) correlated these rocks with ca. 1450–1380 Ma Belt Supergroup strata, suggesting they were the western-most exposures of Belt strata. In Chapter 3 new detrital zircon U–Pb results from the Deer Trail Group were presented. These



results support interpretations like those recently presented by Box et al., (2020), and indicate that the Deer Trail Group is mostly younger than 1300 Ma, and thus not correlative to older Belt Supergroup strata further east. The ages and Lu/Hf values of the detrital zircon components in the Deer Trail Group indicate a mixed Laurentian provenance. Notably, Deer Trail Group strata do not record any non-Laurentian sources, and thus indicate a provenance shift from the sources recorded in the underlying Belt Supergroup. The fine-grained siliciclastic and carbonate lithology, and lack of significant syn-depositional magmatic detrital zircon components in Deer Trail Group strata are consistent with their deposition during a period of regional subsidence within the Priest River region that coincides with purported Nuna breakup at ca. 1300–1200 Ma (Kirscher et al., 2020). These stratigraphic and provenance observations suggest the Deer Trail Group may have been deposited during final rifting and passive margin subsidence associated with Nuna breakup. While the only minimum depositional age constraint on the Deer Trail Group is the maximum depositional age of the overlying, < 760 Ma, Buffalo Hump Formation, it is possible that the Deer Trail Group is much younger than ca. 1200 Ma. For example, it is possible that the Deer Trail Group was instead deposited during the assembly and/or early tenure of Rodinia. However, given the absence of ca. 1200–1000 Ma “Grenville-age” zircon grains in the Deer Trail Group, despite their prevalence in Neoproterozoic strata in the region lends credence to a likely ca. 1300–1200 Ma age for the Deer Trail Group.

During this same general time interval, prior workers identified a tectono-magmatic event that may have terminated deposition of the Belt Supergroup. McMechan and Price (1982) delineated the East Kooteny orogeny as a ca. 1350 Ma period of regional folding, metamorphism, and magmatism within the northern extent (southeastern British Columbia) of the Belt Basin. However, in the more southern extent of the Belt Basin (north central Idaho) Doughty and Chamberlain (1996) instead assert that the East Kooteny “orogeny” was not an orogen but rather a ca. 1380–1330 Ma pulse of bimodal magmatism, basin rifting, and renewed subsidence and sedimentation that shortly preceded the end of the deposition in the Belt Basin.

New results from ca. 1400–1350 Ma metamorphic zircon from a mafic sill within the Clearwater complex (presented in Chapter 2) contain  $\delta^{18}\text{O}$  values consistent with hydrothermal zircon growth. In Chapter 4, metapelitic rocks from the western portion of the Clearwater complex are also shown to contain ca. 1350 Ma and younger ca. 1070 Ma monazite U/Pb ages. Interestingly, within the three studied

samples from the western Clearwater complex, all three-record garnet growth associated within the younger ca. 1070 Ma monazite ages. However, only one sample recorded garnet growth (Zirakparvar et al., 2010; Neishem et al., 2012) associated within the older ca. 1350 Ma monazite ages, suggesting lower pressure (below approximately 6 kbar) metamorphism for this older event. Consequently, the occurrence of hydrothermal zircon growth and apparently high temperature, lower pressure metamorphism, as well as the absence of any coarse-grained strata possibly associated with this event, suggests that the “East Kootenay Orogeny” is more consistent with a period of bimodal magmatism and basin rifting within the Clearwater complex, then tectonic thickening.

### **7.2.3 Neoproterozoic Laurentia and Rodinia breakup.**

The correlation of ca. 1200–1000 Ma “Grenville-age” orogens have been critical in the construction of various Rodinia models (e.g. Dalziel et al., 1991; Karlstrom et al., 2000; Li et al., 1995). Consequently, the absence, or presence of a poorly preserved/recognized, Grenville-age orogen in western Laurentia is a significant constraint that accurate paleogeographic models must account for. Within this work, the metamorphic (Chapter 4) and tectonostratigraphic (Chapters 3 and 5) record of a potential Grenville-age orogen in the Belt Basin region was investigated. We note that no direct record of ca. 1250–800 Ma magmatism in the Belt Basin region was found by this, or any previous studies.

In Chapter 4, existing Lu/Hf garnet geochronology (Zirakparvar et al., 2010; Neishem et al., 2012) and new monazite and apatite U-Pb results and Thermocalc (pressure-temperature) constraints from ca. 1450 Ma Belt Supergroup metapelitic strata within the western portion of Clearwater complex (notably within the Piegan Group of the Belt Supergroup) indicate that Phanerozoic Cordilleran conditions did not exceed ~550–450°C in this region. Consequently, the mineral assemblages within these rocks likely reflect a high-grade, Grenville-age, metamorphic event recorded by ca. 1100–1060 Ma garnet Lu/Hf and monazite U/Pb ages. Thermodynamic modeling of these mineral assemblages indicate they are consistent with pressures of 7.5–5 Kbar, and temperatures of 670–620°C, suggesting burial to a depth of ~22–17 km or greater around along a moderately-high average geothermal gradient of ~36–23°C/km. Elsewhere in the Belt Basin, and I suggest spatially associated with inherited lithospheric-scale faults, similar ca. 1100–1070 Ma metamorphic events are noted

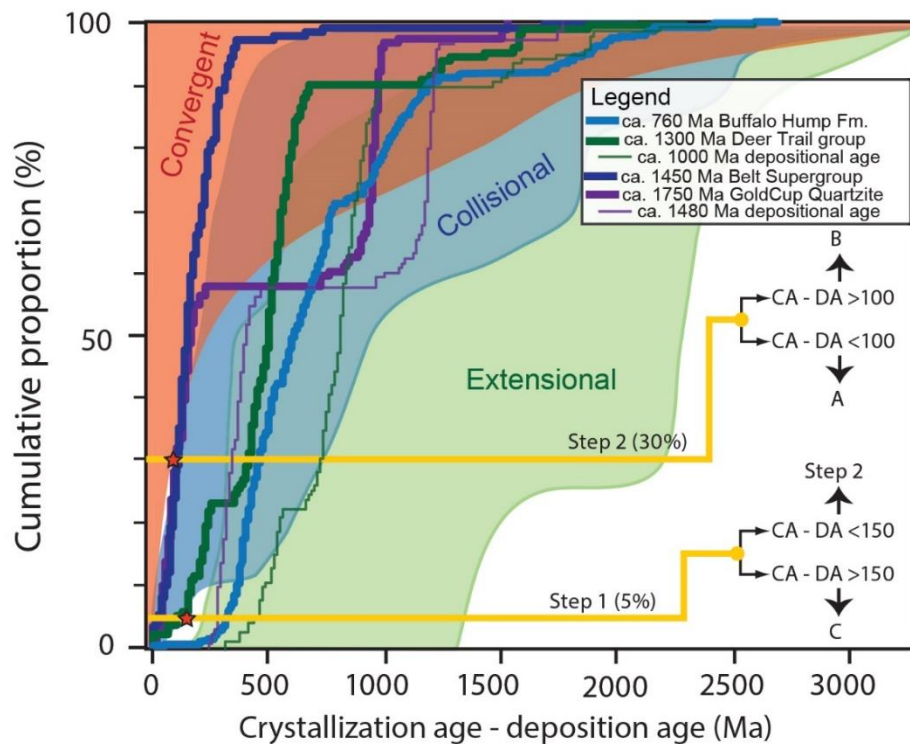
(McFarlane and Pattison, 2000; Aleinikoff et al., 2012; Pattison and Seitz, 2012; Bookstrom et al., 2016).

At ca. 1100 Ma Laurentia was undergoing a period of widespread extension and thermal upwelling. Manifestations of this event are found in the 1110–1085 Ma Midcontinent Rift (e.g. Swanson-Hysell et al., 2019), 1115–1078 Ma Pikes Peak batholith (Guitreau et al., 2016), 1094–1080 Ma Southwestern Laurentian Large Igneous Province (Timmons et al., 2001; Bright et al., 2014), and 1090–1040 Ma Bylot and correlative basins in northern Laurentia (Greenman et al., 2021). Despite the evidence for Laurentian-wide extension and thermal upwelling at this time, it is indeed difficult to account for pure depositional burial of the Piegan Group (of the Belt Supergroup) to a depth ~22–17 km, when the preserved thicknesses of overlying Belt strata are >10 km, and Deer Trail Group strata are ~3.5 km thick. Therefore, the ca. 1110–1060 Ma period of continent-wide thermal upwelling and reactivation of crustal weaknesses, may have indeed combined with localized inversion of the Belt Basin to produce the metamorphism present in the Clearwater complex.

Importantly, as demonstrated in Chapter 5, if there was indeed any Grenville-age localized inversion of the Belt Basin, and the stratigraphic record of this event was preserved, it must be reflected in the Deer Trail Group, whose fine-grained lithology and recycled detrital zircon provenance signature is a better match for an extensional basin than typical convergent or collisional sequences (Fig. 7.3). Alternatively, no strata was deposited during local inversion of the Belt Basin, or it was subsequently eroded, likely prior to Cryogenian rifting.

Previously, the only possibly suspected Grenville-age (ca. 1000 Ma) strata in western Laurentia was the Buffalo Hump Formation. Prior workers had suggested it required a western Grenville-age source based in part on its erroneous correlation to finer-grained Belt Supergroup strata further east (Miller and Whipple, 1989), and that the Buffalo Hump Formation contained ca. 1100 Ma detrital zircon grains (Ross et al., 1992). However, in Chapter 5, the identification of a small, yet significant, ca. 760 Ma detrital zircon population in the Buffalo Hump Formation indicates that it was deposited during Rodinia breakup, not assembly. The presence of large (20+ cm) reworked Deer Trail Group cobbles in the Buffalo Hump Formation, and the absence of the Buffalo Hump strata along with thinned to absent Deer Trail Group strata east of a regional thrust fault (the Jumpoff Joe Fault) suggests that the thrust is a reactivated Neoproterozoic normal fault. Consequently, the Buffalo Hump Formation records

early ca. 760–720 Ma Rodinia fault-controlled rift-basin formation similar to coeval deposits further south in the Uinta, Chuar and Pahrump basins. These results constrain the onset of rifting in the Belt Basin region to after ca. 760 Ma and thus about 70 Myrs later than in the Adelaide Superbasin of the Gawler craton in southeastern Australia (Preiss, 2000; Lloyd et al., 2020). This time difference makes it difficult for the Belt Basin region (western Laurentia) and the Adelaide Superbasin region (Gawler craton) to be direct conjugate margins as is suggested by Rodinia SWEAT models (e.g. Moores et al., 1991).



**Fig. 7.3:** Cumulative proportion curves of the general fields the variation of the difference between the measured crystallization age for a detrital zircon grain, and the depositional age of the succession in which it occurs for convergent, collisional, and extensional basins (from Cawood et al., 2012). Plotted curves show the strata investigated in this work, with the Gold Cup Quartzite in purple, Belt Supergroup in dark blue, Deer Trail Group in green, and Buffalo Hump Formation in light blue. For the Gold Cup Quartzite and Deer Trail Group, whose depositional ages are only broadly constrained, the maximum depositional age approximated by the youngest detrital zircon population is the thicker line, and potential younger minimum depositional ages are plotted as the same color but thinner lines.

Importantly, deposition of the Buffalo Hump Formation during Rodinia rifting, not assembly, along with the accompanying detrital zircon populations in the Buffalo Hump Formation indicates that its detrital zircon components came from eastern/southeastern Laurentia including the Grenville and intervening provinces. This detritus was likely carried via pancontinental fluvial systems that drained the Grenville orogeny as is suggested for other Neoproterozoic western, and northern Laurentia strata that contain similar detrital zircon age-components (e.g. Rainbird et al., 1992;

2017, Yonkee et al., 2014). During the Neoproterozoic–Cambrian Sauk transgression, many stratigraphic sections along the western U.S. Laurentian margin, record a provenance shift (often roughly correlated with the Neoproterozoic/Cambrian boundary, e.g. Linde et al., 2014) involving abundant ca. 1200–1000 Ma detrital zircon grains to their absence. Despite the widespread nature of this provenance shift, a tectonic mechanism and an understanding of how these Laurentian-wide provenance trends relate to Rodinia breakup processes was lacking.

In Chapter 6, a compilation of Stenian (ca. 1200–1000 Ma) zircon (detrital and igneous) U/Pb and Lu/Hf values were compiled and statistically compared. The results indicated that while eastern Laurentian Grenville province sources are the most likely source for Stenian detrital zircon grains in Tonian and Cyrogenian strata (like the Buffalo Hump Formation), more southern Llano Inlier and Wichita province sources are a better match for Stenian detrital zircon grains in Ediacaran strata. The trends also suggest that the eventual loss of far-travelled Stenian detrital zircon age-components progressed from north to south (Idaho to Arizona) from ca. 540 to 510 Ma. Consequently, these new results suggest that the generally north to south opening of the Iapetus Ocean along the eastern margin of Laurentia (near the center of Rodinia) results in the migration of a drainage divide from eastern to southern to the midcontinent of Laurentia, known as the Transcontinental Arch (e.g. Sloss, 1988). Importantly, this tectonic model provides a mechanism for the signification continent-scale shifts in Laurentian sedimentary systems during Rodinia breakup, and further emphasizes that the eastern and western margins of Laurentia not only differed in character during the assembly of Rodinia (see Chapter 5) but that these margins also differed (perhaps influenced by inheritance of older assembly related features) in character during the breakup of Rodinia.

### **7.3 From Laurentia assembly to Rodinia breakup: A revised early tectonic evolution of the Belt Basin region and its importance for global models**

This work provides several important updates to the timing, importance, and correlation of Proterozoic geologic events in the Belt Basin region, some of which were foundational in the proposal of various Precambrian paleogeographic models and thus have importance that ranges well beyond western Laurentia.

Almost thirty years ago, Link et al., (1993) wrote:

*“First, the Neihart Quartzite at the base of the Belt (Schieber, 1989a) may be part of a supermature discontinuous quartz sand body (Freeman and Winston,*

*1987) that blanketed the craton after the Early Proterozoic assembly of Laurentia (about 1,900 to 1,700 Ma; Hoffman, 1989) and before faulting and subsidence of the Belt Basin (about 1,450 Ma). If the Neihart is a genetically distinct unit, it should be removed from the Belt.”*

In 1998, Doughty et al., suggested that the Gold Cup Quartzite in the Priest River complex is a western equivalent to the Neihart Quartzite based on similarity in detrital zircon U/Pb ages. Mueller et al., (2016) presented further arguments for an approximately 200 m.y. “pre-Belt” age for the Neihart Quartzite based on additional detrital zircon age result differences between the Neihart Quartzite and the Prichard Formation (lower Belt Supergroup). In the simplest stratigraphic sense, Walter’s Law of stratigraphic succession (i.e. any vertical progression of facies is the result of a succession of depositional environments that are laterally juxtaposed to each other; Middleton, 1973) dictates a disconformity between the coarse, pebbly fluvial to nearshore Goldcup Quartzite and the overlying lower slope deposits of the basal (member A) Prichard Formation (Cressman, 1985). This relationship is consistent with the tectonized unconformity contact identified by Doughty and Chamberlain (1998) in the field. The additional detrital zircon U/Pb and Lu/Hf data presented in Chapter 3 further support both these findings, and thus I tend to support the suggestion that the Neihart and Gold Cup quartzites both be “removed from the Belt” indicating that the unconformity between lower Belt strata and these pre-Belt quartzites may represent a significant ~300–200 Myr duration. I encourage further work on the physical, temporal and tectonostratigraphic relationships of the allegedly lower Belt equivalents within the Helena embayment (see Mueller et al., 2016). None of the detrital zircon components within these units suggest a non-Laurentian provenance, and consequently the strata in the Belt Basin region do not require Nuna assembly prior to ca. 1750 Ma, and are permissible with models (e.g. Nordsvan et al., 2018) of a later 1650–1600 Ma final Nuna assembly.

It has been long recognized that the source region of the western Belt Basin lay to the west of the basin and was removed by subsequent rifting (Ross et al., 1992). Detrital zircon studies of western Belt Basin strata revealed that this source terrane was primarily comprised of ca. 1700–1500 Ma rocks, including ca. 1610–1490 Ma “North American Magmatic Gap” ages, which are rare in Laurentia, but abundant in southeastern Australia, establishing early ties between these two regions during the Mesoproterozoic (Ross and Villeneuve, 2003). However, identification of other ca.

1590–1550 Ma “North American Magmatic Gap” granitoids in East Antarctica (Goodge et al., 2017), south China (Xu et al., 2019), and within the Belt Basin region itself (e.g. Evans and Fischer, 1986) brought into doubt the uniqueness of the western Belt Basin, southeastern Australia provenance tie. However, currently only the Gawler Craton of southeastern Australia contains ca. 1590–1550 Ma granitoid sources (Reid and Payne, 2017) that match the juvenile Lu/Hf values of the similar-age zircon grains found within the western Belt Basin (Fig. 7.2; see Chapters 2 and 3). Further testing of this relationship by other detrital phases (e.g. rutile, monazite) may yield interesting results into the tectonic setting of this relationship (e.g. Pereira et al., 2020). The similar timing, and proposed genetic relationship between ca. 1580 Ma magmatism in the Priest River region, and coeval magmatism in southeastern Australia, suggest that these two regions may have been adjacent to each other within Nuna by ca. 1580 Ma, consistent with paleomagnetic constraints (see Fig. 7.5; Kirscher et al., 2019).

The absence of juvenile ( $\epsilon_{\text{Hf}} > +5$ ) ca. 1700–1500 Ma detrital zircon grains within ca. 1450 Ma successions on Hainan Island of south China, means that the detrital zircon provenance rationalization for featuring Hainan Island (Xu et al., 2019), and sometimes the entire Cathasia block (e.g. Yao et al., 2017) between the Belt Basin and SE Australia within Nuna is not strongly justified. However, the sample size (21 zircon grains, Xu et al., 2019) from ca. 1450 Ma Hainan Island strata that include Lu/Hf data is relatively small and additional analysis may refine this assertion. Other evidence commonly used to justify the central location of Hainan Island and the Cathaysia block in Nuna, such as correlating ca. 1450 Ma felsic magmatism on Hainan Island with the ca. 1500–1350 Ma Granite-Rhyolite province of southern Laurentia (Li et al., 2008b) is permissible. Magmatism at this time was generally widespread along the margin of Nuna, with coeval magmatism recognized in other cratons like Baltica, Amazonia, and India (e.g. Johansson et al., 2022; Sequeira et al., 2022).

Many of the Paleoproterozoic to early Mesoproterozoic correlations discussed above and in prior chapters of this work were incorporated into early reconstructions of supercontinent Rodinia (e.g. Bell and Jefferson, 1987; Moores et al., 1991; Dalziel, 1991; Hoffman, 1991). Thus, at least some of the evidence for a Neoproterozoic SWEAT-like **Rodinia** connection was, in fact, evidence of a Mesoproterozoic (proto-) SWEAT-like **Nuna** connection. In part, the apparent absence of Young et al.’s (1979) “succession B” or Grenville-age (ca. 1200–1000 Ma) strata between ca. 1480–1380 Ma Belt and < 720 Ma Windermere Supergroup strata in the Belt Basin region (see

Link et al., 1993), led many of these early models to advocate a long-lived relative tectonic relationship between Australia and western Laurentia during the Mesoproterozoic to Neoproterozoic. However, other subsequent models (such as the missing-Link; Li et al., 1995; 2002; 2013) rather advocated for a complete Wilson cycle (e.g. opening and closing of an ocean basin) along western Laurentia during Grenville-time. One line of reasoning to support this model was the alleged presence of “succession B” or Grenville-age (ca. 1200–1000 Ma) orogenesis and foreland basin strata recorded within the Buffalo Hump Formation along the western edge of the Belt Basin. Consequently, this thesis took a critical look at the geologic record of a potential western Grenville-age orogen within the Belt Basin region, which until this work (Chapter 3, 4, and 5) was not vigorously tested.

In Chapter 5, the speculated South China-derived ca. 1100–1000 Ma foreland basin deposit within the Buffalo Hump Formation (Li et al., 1995; 2002; Yao et al., 2017) was investigated. While, the Buffalo Hump Formations’ coarse nature, and conglomeratic horizons did indeed suggest proximal uplift during its deposition, the small yet significant ca. 760 Ma detrital zircon population, and regional stratigraphic relations, indicated that the Buffalo Hump Formation was not an exotically derived Grenville-age orogenic deposit. Instead, the Buffalo Hump Formation is a Laurentian sourced ca. 760–720 Ma rift basin deposit. The unconformably underlying Deer Trail Group strata (see Chapter 3), could possibly be ca. 1200–1000 Ma in age, given that they contain a ca. 1300 Ma youngest detrital zircon population. However, Deer Trail strata lack Grenville-age detritus but otherwise exhibit a clear Laurentian provenance signature. Additionally, their fine-grained siliciclastic and carbonate nature precludes adjacent significant uplift. Thus, if the Deer Trail Group is Grenville in age, they are also inconsistent with a proximal orogen and rather suggest a period of extension within the region.

In Chapter 4, the ca. 1400–1000 Ma garnets reported from the Clearwater complex (Zirakparvar et al., 2010; Neishem et al., 2012) and their potential record of a Grenville-age orogen was also investigated. The new monazite U/Pb results support the occurrence of two main metamorphic events, an older ca. 1380–1330 Ma event, and younger ca. 1100–1060 Ma event. The older event was widely recorded in monazite U/Pb ages but only locally resulted in garnet growth, consistent with a lower pressure, higher temperature event. This event likely relates to similar-age hydrothermal zircon growth (see section 2.5.6) and magmatism within the Belt Basin.



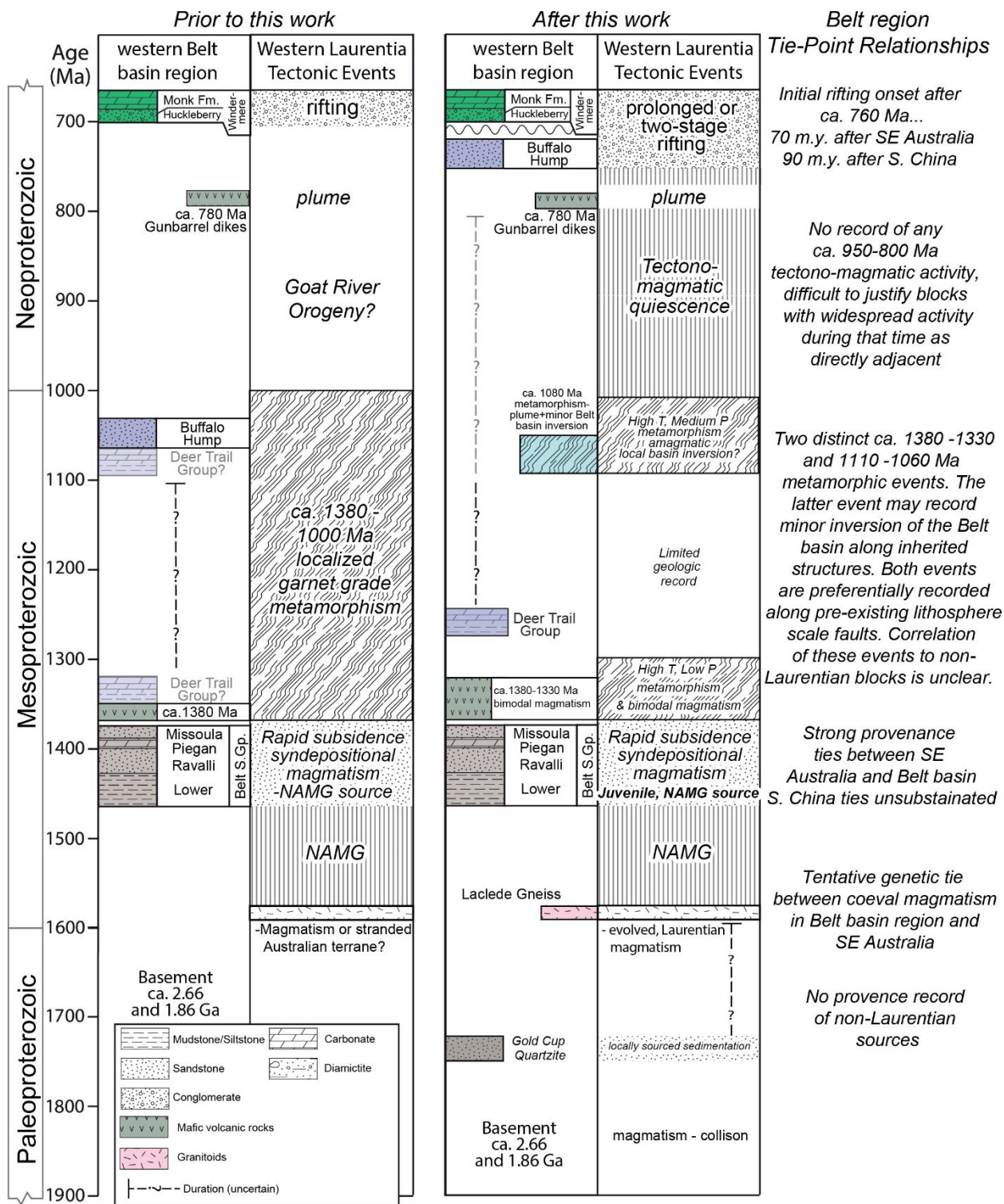
These results are consistent with previous interpretations of ca. 1380–1330 Ma being a period of magmatism and renewed rifting within the Belt Basin (Doughty and Chamberlain, 1996) and do not require orogenic thickening. However, the Pressure-Temperature-time constraints suggest that the younger ca. 1100–1060 Ma event recorded burial of middle Belt Supergroup strata (Piegan Group) to depths of at least ~17–22 km along a moderately-high average geothermal gradient. Elsewhere in Laurentia at this time, similar age geology records mantle/thermal upwelling, intraplate volcanism, reactivation of crustal weaknesses and extension such as within the 1110–1085 Ma Midcontinent Rift (e.g. Swanson-Hysell et al., 2018), 1115–1078 Ma Pikes Peak batholith (Guitreau et al., 2016), 1094–1080 Ma Southwestern Laurentian Large Igneous Province (Timmons et al., 2001; Bright et al., 2014), and 1090–1040 Ma Bylot and correlative basins in northern Laurentia (Greenman et al., 2021; see Fig. 4.2).

Consequently, thermal upwelling and reactivation of crustal weaknesses within the Belt Basin likely contributed to the ca. 1100–1060 Ma metamorphism. However, there is an approximately 6 km discrepancy between the calculated burial depth, and the preserved thickness of overlying units, suggesting either A) a component of tectonic burial, likely through localized inversion of the Belt Basin, and related burial of Belt strata in the Clearwater region or B) miscalculation of Belt Supergroup and perhaps Deer Trail Group stratigraphic thickness and/or subsequent erosion. However, if calculated pressures do reflect tectonic burial, this tectonism was amagmatic, localized, associated with syndepositional Belt Basin structures, and produced very little (to no) preserved sedimentation. Thus, of the multiple working models presented in Chapter 4, transpressional/transensional mechanism, rather than subduction-driven continent-continent collisional orogenic tectonic mechanism may be more likely. Importantly, after the termination of this metamorphic event around ca. 1060 Ma, there is no record of any magmatism and/or tectonism in the Belt Basin region until the intrusion of the Gun Barrel dyke swarm around 780 Ma (Harlan et al., 2003). In the Belt Basin region, recalibration of the Buffalo Hump Formation (see Chapter 5) indicates that ca. 780 Ma dyke intrusion was followed by widespread but localized rift-basin development from ca. 760–720 that correlates to other localized Laurentian basins. Collectively, these “CHUMP-B” basins reflect early Rodinia breakup (e.g. Dehler et al., 2017; Chapter 5). As discussed in Chapters 5 and 6, in the Belt Basin region, intermittent Cryogenian–Cambrian sedimentation was likely influenced by

local rift structure/timing as well as Laurentian-scale drainage divide migrations associated with circum-Laurentia rifting. Importantly, Neoproterozoic–Cambrian strata in the Belt Basin region record final rifting and breakup in the late Ediacaran–early Cambrian (Lund et al., 2010; Lund and Cheney, 2016; Brennan et al., 2020; 2022).

This revised geologic record (Fig. 7.4) is difficult to reconcile with a SWEAT-like (e.g. Moores, 1991; Merdith et al., 2017) configuration which places southeastern Australia, and its record of mantle plume and rifting activity as early as 830 Ma, as a direction conjugate to Belt region within Rodinia. A Missing-Link configuration (Li et al., 1995; 2013) which was initially proposed to alleviate some of the same discrepancies with a SWEAT configured Rodinia, highlighted above, instead places South China, and its significant ca. 1300–900 Ma Grenville-age orogenesis (Li et al., 1995; 2007), widespread ca. 950–830 Ma volcanism (Yao et al., 2019), and record of rifting as early as 850 Ma (Li et al., 2008) directly conjugate to the Belt region within Rodinia. Thus, a Missing-Link configuration, with South China adjacent to the Belt Basin region of Laurentia near the center of Rodinia, is also not a perfect fit for allegedly adjacent regional geologic histories. Recently, an external or peripheral location of South China within Rodinia is often advocated due to South China's Neoproterozoic tectonostratigraphic similarity with India (e.g. Cawood et al., 2013) and some paleomagnetic constraints (Park et al., 2021; Chang et al., 2022).

As a result, the regional geologic constraints discussed and redefined throughout this thesis highlights issues with both of the most commonly applied Rodinia models, and echos the ethos of recent paleomagnetic reconstructions that also indicate a tight-fitting SWEAT or Missing-Link configuration may be untenable within Rodinia (Eyster et al., 2020; Park et al., 2021; Swanson-Hysell, 2021; Chang et al., 2022). Consequently, perhaps neither southeastern Australia, nor South China were directly adjacent to the Belt Basin region within Rodinia. As speculated in Chapter 5, widespread, perhaps multiphase ca. 760–540 Ma rifting and passive margin sequences along most of western Laurentia may instead reflect the rifting of other smaller blocks (recent work suggests Tasmania or Tarim may be suitable candidates) or yet-to-be identified ribbon continents (e.g. Mulder et al., 2020; Jing et al., 2021), rather than a single protracted rifting event between Laurentia and eastern Australia or Laurentia and South China (see Fig. 7.5).



**Fig. 7.4:** Left: Time-space diagram for the Paleo–Neoproterozoic geology and corresponding tectonic events of the western Belt basin region (northeastern Washington and northern Idaho, USA), and major western Laurentian tectonic events (ca. 2018) prior to the onset of this work. Center: The revised Time-space diagram for the western Belt basin region resulting from the data and conclusions presented in this work. Right: Summary of the tie-point relationships of the Belt basin region and their importance for paleogeographic tectonic reconstructions.



Collectively, the redefined tectonic history of the Belt Basin region strengthens models of a (proto-) SWEAT configured Nuna supercontinent from ca. 1600–1300 Ma (Kirscher et al., 2020) but identified tectonostratigraphic issues with these same regions as directly conjugate rift margins within subsequent supercontinent Rodinia (Fig. 7.5). I also emphasize that some of the original SWEAT correlations made in support of early Rodinia models (Moores, 1991; Dalziel, 1991; Hoffman, 1991), such as western derived detritus in the Mesoproterozoic Belt Basin, instead support a (proto) SWEAT-like relationship during Nuna-time. On the other hand, several of the foundational assertions that the Missing-Link model relied upon, such as the shared provenance between Hainan Island and the Belt Basin and Buffalo Hump Formation, (Li et al., 1995; 2008; Yao et al., 2017), considering new data are worth re-evaluating. However, it is worth noting that a Missing-Link configuration, notably the significant ca. 1300–900 Ma Grenville-age orogenesis and Neoproterozoic plume record in South China (e.g. Li et al., 1995; 2008), does perhaps provide the most compatible geologic record to reconcile the aforementioned differences with a SWEAT-like configured Rodinia.

Consequently, while the value and importance of global tectonic models for understanding and predicting fundamental earth processes is undeniable, it is clear that after more than 30 years of focused research, a clear consensus on Proterozoic paleogeography is still lacking. Thus, I respectfully submit that the work of this thesis espouses the foundational tenet that scientific progress requires that all viable models remain open to observation-driven modification and refinement. With anticipation, openness, and humility, I look forward to new observations testing, refining and perhaps refuting the interpretations I present here.

## 7.4 References

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**Appendix A: Statements of Co-Authorship  
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## Appendix A1: Statement of Co-Authorship for Chapter 2

Chapter 2: Closing the “North American Magmatic” Gap: Crustal evolution of the Clearwater Block from multi-isotope and trace element zircon data *published in Precambrian Research*

	Conceptualization and Design	Data Acquisition and Method	Data processing and Visualization	Interpretation and discussion	Total contribution (%)
<b>Co-author 1</b> Daniel T. Brennan	75	75	80	70	75
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<b>Co-author 2</b> Paul K. Link	15	-	5	10	7.5
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<b>Co-author 3</b> Zheng-Xiang Li	10	-	5	10	6.25
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<b>Co-author 4</b> Laure Martin	-	15	5	-	5
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<b>Co-author 5</b> Tim E. Johnson	-	-	-	10	2.5
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<b>Co-author 6</b> Noreen J. Evans	-	5	5	-	2.5
Co-author 6 acknowledgement: I acknowledge that these represent my contribution to the above research output.			Signed:		
<b>Co-author 7</b> Jiangyu Li	-	5	-	-	1.25
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### Appendix A2: Statement of Co-Authorship for Chapter 3

Chapter 3: Detrital zircon U-Pb and Hf signatures of Paleo-Mesoproterozoic strata in the Priest River region, northwestern USA: A record of Laurentia assembly and Nuna Tenure *published in Precambrian Research*

	Conceptualization and Design	Data Acquisition and Method	Data processing and Visualization	Interpretation and discussion	Total contribution (%)
<b>Co-author 1</b> Daniel T. Brennan	50	85	85	60	70
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<b>Co-author 2</b> J. Brian Mahoney	20	10	10	10	12.5
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<b>Co-author 3</b> Zheng-Xiang Li	20	-	-	15	8.75
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<b>Co-author 4</b> Paul K. Link	10	-	-	10	5
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<b>Co-author 5</b> Noreen J. Evans	-	5	5	-	2.5
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<b>Co-author 6</b> Tim E. Johnson	-	-	-	5	1.25
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### Appendix A3: Statement of Co-Authorship for Chapter 5

Chapter 5: Recalibrating Rodinian rifting in the northwestern United States *published in Geology*

	Conceptualization and Design	Data Acquisition and Method	Data processing and Visualization	Interpretation and discussion	Total contribution (%)
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<b>Co-author 2</b> Zheng-Xiang Li	20	-	-	15	8.75
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<b>Co-author 3</b> Kai Rankenburg	5	15	10	-	7.5
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<b>Co-author 4</b> Noreen Evans	-	15	10	-	6.25
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<b>Co-author 5</b> Paul K. Link	5	-	-	10	3.75
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<b>Co-author 6</b> Adam Nordsvan	-	-	5	5	2.5
Co-author 6 acknowledgement: I acknowledge that these represent my contribution to the above research output. Signed:					
<b>Co-author 7</b> Christopher L. Kirkland	-	5	5	-	2.5
Co-author 7 acknowledgement: I acknowledge that these represent my contribution to the above research output. Signed:					
<b>Co-author 8</b> J. Brian Mahoney	-	-	-	5	1.25
Co-author 8 acknowledgement: I acknowledge that these represent my contribution to the above research output. Signed:					
<b>Co-author 9</b> Tim Johnson	-	-	-	5	1.25
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<b>Co-author 10</b> Bradley J. McDonald	-	5	-	-	1.25
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### Appendix A4: Statement of Co-Authorship for Chapter 6

Chapter 6: A tectonic model for the Transcontinental Arch: Progressive migration of a Laurentian drainage divide during the Neoproterozoic–Cambrian Sauk Transgression *published in Terra Nova*

	Conceptualization and Design	Data Acquisition and Method	Data processing and Visualization	Interpretation and discussion	Total contribution (%)
<b>Co-author 1</b> Daniel T. Brennan	50	80	80	50	65
Co-author 1 acknowledgement: I acknowledge that these represent my contribution to the above research output.			Signed:		
<b>Co-author 2</b> Ross N. Mitchell	30	10	15	15	17.5
Co-author 2 acknowledgement: I acknowledge that these represent my contribution to the above research output.			Signed:		
<b>Co-author 3</b> Christopher J. Spencer	10	10	5	15	10
Co-author 3 acknowledgement: I acknowledge that these represent my contribution to the above research output.			Signed:		
<b>Co-author 4</b> J. Brendan Murphy	10	-	10	10	7.5
Co-author 4 acknowledgement: I acknowledge that these represent my contribution to the above research output.			Signed:		
<b>Co-author 5</b> Zheng-Xiang Li	-	-	-	10	2.5
Co-author 5 acknowledgement: I acknowledge that these represent my contribution to the above research output.			Signed:		
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**Appendix B: Conference Abstracts presented during  
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## **An ancient Australian-American Alliance: a multi-isotope investigation of western N. America relations within Proterozoic supercontinents**

**Daniel T. Brennan<sup>1</sup>, Zheng-Xiang Li<sup>1</sup>, Paul K. Link<sup>2</sup>, Tim Johnson<sup>3</sup>, Noreen Evans<sup>4</sup>**

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Along the west-central margin of Laurentia (North America), within the Priest River and Clearwater metamorphic complexes, rare exposures of “North American Magmatic Gap” (NAMG, ca. 1.61-1.49 Ga) crystalline basement rocks of the Clearwater Block are present. Overlying these crystalline basement rocks are ca. 1.47–1.38 Ga Belt Supergroup strata that also contain NAMG-age detrital zircon grains. This unique igneous and detrital record of NAMG-magmatism stands in contrast to other regions of North America that generally lack a record of any ca. 1.61-1.49 Ga tectono-magmatic activity. These differences led some early researchers to speculate that Proterozoic supercontinent events may have stranded exotic, perhaps Australia associated basement terranes within the Clearwater Block and that Australia (or perhaps Antarctica or South China) may have supplied the NAMG zircon grains within the overlying Belt Supergroup. If true, the rocks that now constitute some of the most scenic mountains in the northwestern United States were actually once part of Australia. Here, we report new U-Pb, Lu-Hf and O isotope data from Neoproterozoic and Proterozoic igneous and sedimentary rocks within the Priest River and Clearwater regions of western N. America. Our isotopic results indicate that ca. 1.58 Ga “NAMG” magmatism within the Clearwater block records mantle-like  $\delta^{18}\text{O}$  values but subchondritic ( $-5.5$  to  $-9.7$ )  $\epsilon_{\text{Hf}}$  values in zircon, suggesting (perhaps plume-driven) reworking of the Neoproterozoic country rock. The overlying Belt Supergroup strata contain similar-age detrital zircon grains but with more juvenile suprachondritic (positive)  $\epsilon_{\text{Hf}}$  values that are similar to igneous sources in southeastern Australia. Collectively, these zircon results suggest that the Clearwater Block does have a North American origin, and that detrital zircon grains in the overlying sedimentary rocks of the Belt Supergroup record non-North American sources consistent with rocks in southeastern Australia. Consequently, these new results lend further support for an ancient Australian-American geological alliance.



## Appendix B2: 2021 Australian Earth Sciences Convention Abstract

which point to a global increase in thermal gradients that intersected granite genesis. We suggest these changes occurred as the secular cooling of the mantle and crust was reversed by a net increase in the spatial extent of continental crust between 2–1.8 Ga, resulting in thermal insulation of the mantle. The following 1.2 billion years on Earth was dominated by a warm, insulated mantle and crust, maintained by stable continental volumes, which eventually cooled to allow the second emergence and widespread preservation of eclogites from ca 0.8 Ga until present. While novel, this idea combines unrelated global petrological and geochemical datasets to explore the sensitivity of switches in the thermal evolution of the solid Earth.

### When did Australia's Cratons come together?

Gorczyk, Weronika, Tyler, Ian, Aitken, Alan, & Kohanpour, Fariba

*Centre of Exploration Targeting, School of Earth Science, University of Western Australia, Perth, Australia*

Assembly of Australian cratons as part of Nuna assembly has been a subject of debates for decades. Especially within Australian geoscience community the timing and style of Western Australia Craton (WAC) with Northern Australian Craton (NAC) collision causes a lot of controversy. The dispute arises mostly due to sparsity of data available. As the Proterozoic knowledge of Australian cratons, as well as others grew new models for Nuna assembly were proposed, and the literature became overcrowded with variable models, based on localised and limited data.

Here, without presenting any new data, an attempt is made to analyse existing models in an unbiased style, questioning and correlating all the tectonic and sedimentation events across WAC, NAC and SAC (Southern Australia Craton). The position and interactions with Laurentia and North China – which are believed to be proximal to Australia at the time of paleo-meso Proterozoic, as also considered.

To achieve this task, plate reconstruction software (GPlates) is used. Publicly available geological data that describe tectonic and sedimentary events affecting WAC, NAC and SAC, as well as paleomagnetic data to be taken into account to support or contradict conceptual models. The immense advantage of this approach is continuous space and time visual representation of the plate interactions and occurrence of events.

Three (with variations) time models of WAC and NAC collision are shown with different subduction polarity: (1) 1800–1765 Ma, (2) 1590–1550 Ma, (3) ca 1300 Ma. Essentially, in the first model one can correlate all the tectonic events across WAC and NAC and SAC with one another post-collision, but spatial problem arises between the cratons and events that follow. In the second model the collision of WAC and NAC can be correlated with metamorphic and magmatic events in Arunta region, as well as in Mt Isa, but does not allow for correlation of previous events across WAC and NAC. The third model with subduction under WAC combines the tectonic evolution of Paterson region, Wankanki Arc and Stage I of Albany Fraser in a very elegant way, but again keeps WAC on a separate plate prior 1300 Ma and does not allow for correlation of events across the cratons of with similar styles and

timings. Pros and cons for all models will be presented, and the verdict will be left to you.

### Detrital zircon record of Proterozoic strata in the Priest River region of western Laurentia: Evaluating “SWEAT” relationships for supercontinents Nuna and Rodinia

Brennan, Daniel T.<sup>1</sup>, Li, Zheng-Xiang<sup>1</sup>, Link, Paul K.<sup>2</sup>, & Johnson, Tim<sup>3</sup>

<sup>1</sup>*Earth Dynamics Research Group, School of Earth and Planetary Sciences, Curtin University, WA, Australia;* <sup>2</sup>*Department of Geosciences, Idaho State University, Pocatello, ID, USA;* <sup>3</sup>*School of Earth and Planetary Sciences, The Institute for Geoscience Research (TIGeR), Curtin University, WA, Australia*

Correlation of rocks across purportedly paired margins, such as Proterozoic strata (notably the Belt-Purcell and Windermere Supergroups) of western Laurentia with coeval rocks and/or magmatic sources in and around the Gawler Craton, have long been used as a key piercing point for SWEAT-like reconstructions of supercontinents Nuna and Rodinia. Here we evaluate the nature and timing of the proposed correlations through U–Pb and Lu–Hf analysis of detrital zircon (DZ) from the Proterozoic Gold Cup Quartzite, Belt-Purcell Supergroup, Deer Trail Group, and Buffalo Hump Formation of the Priest River region, northwestern USA.

The <1.7 Ga Gold Cup quartzite contains mostly ca 2.6 and 1.8 Ga DZ grains, indicating it is likely a western equivalent of the Neihart Formation. Lu–Hf values from these grains suggest that the younger ca 1.8 Ga population ( $\epsilon_{\text{Hf}_t} = -9$  to  $-3$ ) resulted from a reworking event on the ca 2.6 Ga crust involving juvenile mantle input ( $\epsilon_{\text{Hf}_t} = -2$  to 4). This is consistent with the sediments being sourced from proximal Neoproterozoic Laurentian terranes such as the Clearwater/Medicine Hat block, that were intruded by Paleoproterozoic magmatism associated with the collision of the Wyoming and Medicine Hat blocks. Thus, these units do not require a SWEAT configuration (or the existence of Nuna) at ca 1.7 Ga. In the overlying western (ca 1.48–1.37 Ga) Belt Supergroup units, significant juvenile ( $\epsilon_{\text{Hf}_t} = 2$  to 8) ca 1.6 Ga DZ grains are present. These grains fall within the North American Magmatic Gap and likely indicate provenance from the Gawler Craton, supporting a proto-SWEAT configuration for Nuna during ca 1.5–1.4 Ga as in most Nuna reconstructions. The overlying <1.3 Ga, fine-grained and carbonate Deer Trail Group is interpreted as a passive margin succession and contains mostly ca 1.9–1.65 Ga DZ grains with a wide range of Lu–Hf values ( $\epsilon_{\text{Hf}_t} = -6$  to 9), notably ca 1.6 Ga DZ grains are absent. This provenance shift could be indicative of Nuna breakup, removal of the Gawler Craton from its Nuna position along western Laurentia, and a southwestern Laurentia provenance or recycling from underlying rocks of the Lemhi group of the Belt-Purcell Supergroup.

Coarse, locally conglomeratic, rocks of the Buffalo Hump Formation unconformably overly Deer Trail group strata. Prior small-n DZ study of the Buffalo Hump Formation identified a ca 1.1 Ga youngest DZ population, which was suggested to record deposition at ca 1.0 Ga during Rodinia amalgamation. However, our large-n study of the Buffalo Hump Formation

identified for the first time a minor (~1%) yet significant ca 760 Ma DZ population, which constrains the maximum age of deposition. These geochronology results redefine the onset of Rodinia rift-related sedimentation to after ca 760 Ma in this region. Additionally, the Buffalo Hump Formation lacks any ca 900–790 Ma DZ grains. Such a DZ age-spectrum, and inferred rift history, is difficult to reconcile with an immediate neighbourhood between Laurentia and Australia in Rodinia as the latter had an earlier start of continental rifting (with ca 830–750 Ma rifting and syn-rift magmatism).

### Review of SHRIMP zircon ages for the Eastern Succession of the Mount Isa and Etheridge Provinces and their provenances

Withnall, Ian

*Geological Survey of Queensland, Brisbane, Australia*

The migration of zircon geochronology data collected by Geoscience Australia (GA) and Geological Survey of Queensland (GSQ) from the Mount Isa Province into the Online Geochron Delivery System, an important repository maintained by GA, provided an opportunity to review the data and replot it using a consistent approach. This included data for which only preliminary plots of had been available to GSQ and never published.

The review highlighted that the main magmatic events that would have contributed zircon to the Eastern Succession sedimentary rocks occurred at 1850–1870 Ma, 1790–1800 Ma, 1780 Ma, 1760 Ma, 1735–1745 Ma, 1725 Ma, 1705–1715 Ma and 1670–1680 Ma and volumetrically smaller events at 1770 Ma, 1755 Ma, 1655–1660 Ma and 1650 Ma.

The Soldiers Cap Group in the easternmost part of the Mount Isa Province and extending under cover to the east is younger than most of the eastern succession. It consists of Llewellyn Formation, Mount Norna Quartzite and Toole Creek Volcanics in ascending stratigraphic order. The Kuridala Group comprises the Starcross Formation and Hampden Slate.

Samples of the two lowermost units of the Soldiers Cap Group and Starcross Formation have similar maximum depositional ages. A closer comparison has been made of their respective provenances by pooling analyses for units in each group. These provenances are similar, indicating a minor, very old source around the Archean–Proterozoic boundary and then almost none up to ca 1900 Ma (the Barramundi Orogeny). Except for minor components from the Kalkadoon–Leichhardt basement (1850–1870 Ma) and Argylla Formation (1780 Ma), by far the major sources appear to be the Wonga–Burstall–Gin Creek plutonic suites at ca 1740 Ma and Fiery Creek Volcanics or Weberra Granite at ca 1710 Ma. They also both have a significant younger component (slightly older in the Soldiers Cap Group at ca 1685 Ma, and ca 1675 Ma in the Starcross Formation). Pooling analyses from the Hampden Slate indicates that apart from the youngest component being ca 1655 Ma, other components are almost identical to those in the Starcross Formation.

By contrast the provenance of the Toole Creek Volcanics is dissimilar to the other units. It shows an ca 1795 Ma, ca 1850 Ma, and ca 2680 Ma.

Comparing the provenance spectra of the lower part of

the Soldiers Cap and Kuridala Groups with those of the lower part of the Etheridge Group in the Etheridge Province (Georgetown region) suggests that they were probably deposited at about the same time, but the provenance patterns are strikingly different. The Etheridge Group shows a large Archean component as well as almost continuous spread of data points throughout the Paleoproterozoic including peaks around 1900–2000 Ma. This dissimilarity has been cited as evidence that the Georgetown rocks were not distal to Mount Isa and were part of Laurentia until welded to the Australian craton during the assembly of Nuna. The provenance of the upper part of the Etheridge Group, however, is like that of the Toole Creek Volcanics.

### Partial melting, granulites, retrogression and their control on late orogenic exhumation processes

Kenki-Tok, Bénédicte<sup>1,2</sup>, Rey, Patrice F.<sup>2</sup>, & Arcay, Diane<sup>1</sup>

<sup>1</sup>*Géosciences Montpellier, Université de Montpellier, CNRS, 34095 Montpellier Cedex 5, France;* <sup>2</sup>*Earthbyte Research Group, School of Geosciences, University of Sydney, NSW 2006, Sydney, Australia*

Orogenesis drives the differentiation of the continental crust through metamorphic and magmatic processes, the exhumation of deep metamorphic terranes and the concomitant formation of sedimentary basins. A major consequence of prograde metamorphism following a typical orogenic thermal gradient is the dehydration and partial melting of buried rocks leading to the formation of migmatites and granulites. Partial melting and granulitisation are often intertwined and primarily linked to the availability of fluids. Here, we consider the thermal and mechanical consequences of coupled partial melting, granulitisation and strain-rate dependent retrogression during the orogenic cycle, in particular during the recovery phase when the crust's thickness and geotherm re-equilibrate. We explore through 2D thermo-mechanical modelling how the interplay between mechanical weakening due to partial melting and mechanical strengthening due to granulitisation impacts the formation and preservation of crustal roots, the exhumation of the partially molten crust in gneiss domes, the formation of HT/UHT terranes and the partitioning of deformation through the crust.

Our results show that the survival of granulites, which strengthen the lower crust and decrease its capacity to flow under gravitational stresses, impedes the formation of migmatite-cored gneiss domes, and controls the formation and preservation of thick and strong granulitic roots. These are strong enough to stay immune to gravitational stresses and persist over hundreds of million years. These can be actually compared with stable intracontinental regions where the presence of localized crustal roots explains the remarkable variability – from 25 to 65 km – of crustal thickness. Finally, our results highlight the importance of an elevated radiogenic heat production in the upper crust in order to form the long-lived HT/UHT terranes often resulting from supercontinents amalgamation. Our experimental results explain as well why some ancient orogenic domains expose at the Earth's surface dominantly granulitic terranes (e.g., South India, Sri

## Appendix B3: International Geoscience Programme 648 Virtual Seminar Series Abstract

### SUPERCONTINENT CYCLES & GLOBAL GEODYNAMICS

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#### IGCP 648 VIRTUAL SEMINAR SERIES


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- Series 3**
- Series 4

Home > IGCP 648 Virtual Seminar Series > Series 3

## Series 3

### Series 3, Seminar 1 - Daniel Brennan

Reconciling paleogeographic reconstructions with the record of Rodinian rifting in the northwestern United States



#### Speaker bio

Daniel Brennan is a PhD student with the Earth Dynamics Research Group at Curtin University in Perth (Australia). Dan's current focus is on the Neoproterozoic regional geologic record of the northwestern U.S., and its implications for global paleogeographic tectonic models. While Dan's published research has consisted primarily of geologic mapping and tectonostratigraphic (U–Pb, Lu–Hf) provenance studies, his current research is branching into more diverse isotope geochemistry and phase equilibrium modeling. Dan also enjoys teaching and filled a temporary (1 semester) associate lecturer position during his PhD where he taught Physical Geology and Geologic Mapping. Prior to undertaking the PhD, Dan completed a B.Sc. at University of Wisconsin–Eau Claire (2016), and an M.Sc. at Idaho State University (2018). At Idaho State, Dan's award-winning geologic mapping (2019 USGS Student Geologic Map 1<sup>st</sup> place winner) redefined a sequence of Neoproterozoic and lower Paleozoic strata. Dan is scheduled to finish his PhD in January 2022.

#### Abstract

Supercontinent Rodinia assembled through widespread (Grenvillian) orogenic events between 1300 Ma and 900 Ma, and much like its assembly experienced a diachronous break up, perhaps recorded by localized rifting as early as 825 Ma. The western margin of Laurentia (North America) occupied a critical, central location within Rodinia. Despite its important location and its relatively well-studied nature, the evolution of these processes in western Laurentia and the identity of its conjugate margins is a subject of long-standing debate. Here, we present results from the northwestern United States (Washington State), including the Buffalo Hump Formation. The Buffalo Hump Formation unconformably overlies ca. 1.3 Ga strata, and was previously interpreted by some to be a ca. 1.0 Ga "Grenvillian" deposit formed during Rodinia assembly. Our study of ~2000 detrital zircon grains from the Buffalo Hump Formation identified for the first time a minor (~1%) yet significant ca. 760 Ma population, which constrains the maximum age of deposition. This indicates an absence of Grenville-age sedimentation and a ca. 760 Ma earliest onset of Neoproterozoic rifting in this region which must be accounted for in paleogeographic tectonic reconstructions.

## Appendix B4: 2019 Geological Society of America Annual Meeting



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### GSA Annual Meeting in Phoenix, Arizona, USA - 2019

Paper No. 127-4

Presentation Time: 9:00 AM-6:30 PM

DETRITAL ZIRCON U/PB AND LU/HF ANALYSES OF THE BUFFALO HUMP FORMATION AND UNDERLYING DEER TRAIL GROUP: INSIGHTS INTO THE PROTEROZOIC TECTONOSTRATIGRAPHY

**BRENNAN, Daniel T.**<sup>1</sup>, LI, Zheng-Xiang<sup>2</sup>, NORDSVAN, Adam<sup>2</sup>, LINK, Paul K.<sup>3</sup>, MAHONEY, J. Brian<sup>4</sup>, YAO, Weihau<sup>2</sup> and PARKER, Stuart D.<sup>5</sup>, (1)Department of Applied Geology, University of Wisconsin - Eau Claire, Eau Claire, WI 54703, (2)Department of Applied Geology, Curtin University of Technology, Department of Applied Geology, Perth, Australia, (3)Geosciences, Idaho State University, Pocatello, ID 83209, (4)Geology, University of Wisconsin - Eau Claire, Eau Claire, WI 54703, (5)Idaho State University, Pocatello, ID 83209

Significant uncertainties regarding the Proterozoic tectonic history of the western margin of Laurentia hampers our understanding of the timing, geometry, and the occurrence of a complete Wilson cycle along the western margin of Laurentia during the Nuna to Rodinia transition has been suggested, substantiating the sedimentary, metamorphic, and structural record of the region.

The Deer Trail Group and the overlying Buffalo Hump Formation constitute a ~3km thick siliciclastic and dolomitic sequence that are typically thought to be Belt Supergroup. Early work suggested a possible exotic western provenance for some of the units. Here we present new detrital zircon data that revise the provenance of the predominantly argillaceous and dolomitic Deer Trail Group (Togo Fm.) has a maximum depositional age (MDA) of ~1365 Ma, but consist mostly of a ~1.6 to 1.0 Ga younger than, but shares similar sources with, the upper Belt Supergroup equivalent strata in the Lemhi sub-basin. Up-section, detrital zircon spectra from the Deer Trail Group show a dominant Stenian (1.2–1.0 Ga) population that shows juvenile Lu-Hf isotopic values ( $\epsilon_{\text{Hf}}$  of +2 to +12). Several tectonic models propose the existence of a proximal, non-Laurentian source for this detritus at ca. 1.0 Ga. However, nearby (<780 Ma) Windermere Supergroup zircon populations that are interpreted to have been sourced from distal Laurentian Grenville provinces.

This new data corroborates Box et al.'s (2019; GSA Cordillera abstract) that the Buffalo Hump Formation likely unconformably overlies the Deer Trail Group, including paleocurrent analysis, is required to evaluate their potential for constraining the late Proterozoic tectonic evolution of western Laurentia and its provenance.

Session No. 127--Booth# 324

T50. Structure and Tectonic Studies, from Outcrop to Supercontinent (Posters): In Honor of Ian Dalziel  
Monday, 23 September 2019: 9:00 AM-6:30 PM

Hall AB, North Building (Phoenix Convention Center)

Geological Society of America *Abstracts with Programs*. Vol. 51, No. 5  
doi: 10.1130/abs/2019AM-338099

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[Back to: T50. Structure and Tectonic Studies, from Outcrop to Supercontinent \(Posters\): In Honor of Ian Dalziel](#)

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## **Appendix C: Non Peer-Reviewed Science Outreach Articles**



**Appendix C1: A Short Hike on the Edge of Ancient North America: Out There  
Outdoors Magazine, 2021**



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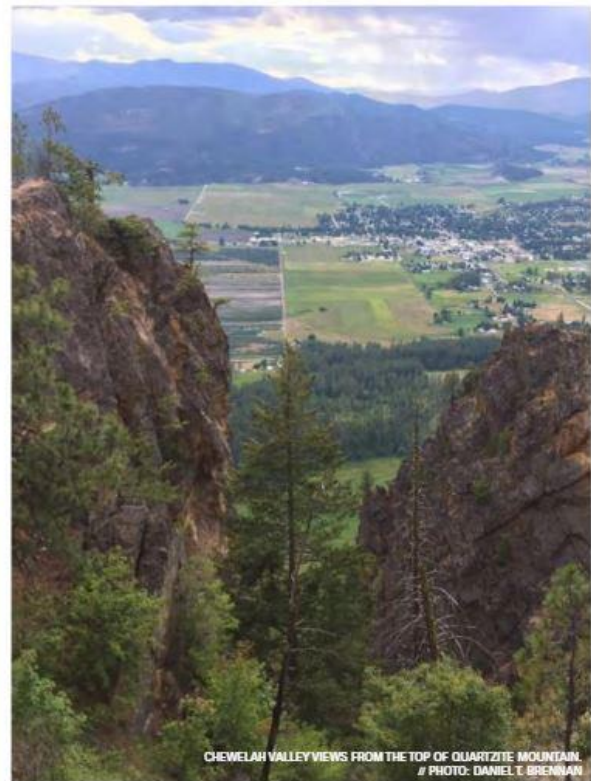
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**HIKING**

**Quartzite Mountain**  
A Short Hike on the Edge of  
Ancient North America

By Daniel T. Brennan



CHEWELAH VALLEY VIEWS FROM THE TOP OF QUARTZITE MOUNTAIN.  
# PHOTO: DANIEL T. BRENNAN

**THE CHEWELAH VALLEY**, like any good valley, is flanked on both sides by mountain ranges. To the east is the Selkirk Range and to the west are the Huckleberry Mountains. However, these mountains are relatively young geologically—we are talking only tens of millions of years.

**GEOLOGIC HISTORY OF THE CHEWELAH VALLEY**

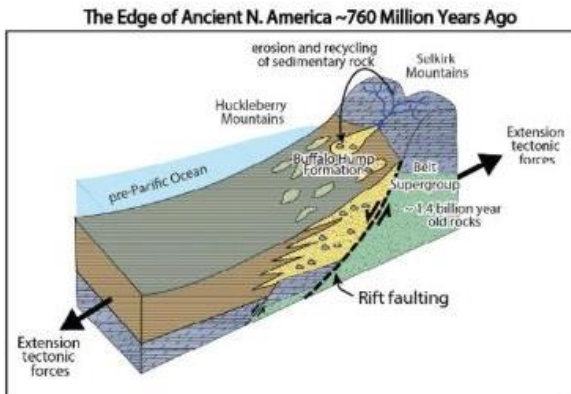
Long before the rocks that make up the Selkirks and Huckleberry Range were uplifted into mountains, geologists have suggested western North America was connected to an unidentified continent, perhaps Australia, around 1 billion years ago. Scientists consider ancient configurations such as this one as a "supercontinent," this one being called "Rodinia," which comes from the Russian word "to give birth" since geologists once thought all subsequent continents spawned

from the breakup of Rodinia. You may be more familiar with the most recent supercontinent called Pangea that existed approximately 300 million years ago.

Around 780 million years ago, tectonic forces broke apart supercontinent Rodinia, and a new pre-Pacific Ocean formed. Since the rocks that make up the western portions of Washington State and Oregon are significantly younger than 780 million years old, this suggests that the ancient edge of North America, that formed during breakup of supercontinent Rodinia, lies unidentified somewhere in the mountains of eastern Washington.

Research, by myself and other colleagues that was recently published in the peer-reviewed scientific journal "Geology," suggests that this supercontinent breakup process occurred directly beneath the Chewelah Valley about 760 million years ago. All it takes is a short but steep hike up Quartzite





Mountain, just east of Chewelah, to get an awesome view from what was once the edge of this ancient world.

**SOLVING AN ANCIENT GEOLOGIC MYSTERY**

On Quartzite Mountain, the oldest sedimentary rocks are called the Belt Supergroup and are made of silt, sand, and limestone that were deposited approximately 1.4 billion years ago, likely within a huge ancient lake in the center of supercontinent Rodinia. As you start the steep incline near the Quartzite Mountain trailhead, these are the poorly exposed rocks you are hiking over. However, at the top of Quartzite Mountain, orange blocky outcrops of hard quartzite rocks create a striking cliff. These rocks are only about 540 million years old, indicating a complete absence of the approximately 780-600 million year-old rocks that are thought to have been deposited when supercontinent Rodinia was breaking apart.

However, west across the valley in the Huckleberry Mountains, beneath the same approximately 540-million-year-old quartzite rocks, very different rock layers, or stratigraphy, is present. These different rock layers include small quartzite boulders in a unit called the Buffalo Hump Formation. The nature of the sand and silt layers in the Buffalo Hump indicated to us that it was likely deposited in ancient river channels, deltas, or a shallow basin. By measuring uranium and lead isotopes of certain minerals in these different rock layers, our geologic study revealed that an ancient river deposited these small boulders approximately 760 million years ago, and the boulders were sourced from the approximately 1.3-billion-year-old rocks exposed across the valley, near the top of Quartzite Mountain.

This suggests uplift and erosion of the eastern side of the modern Chewelah valley, and subsidence of the western side sometime

around 760 million years ago. We proposed that the simplest way to explain the ancient topography suggested by these rocks is the existence of an ancient fault, or fracture in the earth's crust, underneath the Chewelah Valley. This fault likely lowered the western portion of the valley and uplifted the eastern portion. Such faults commonly form when continents are pulled apart in a process known as rifting. All of this suggests that the edge of ancient North America was forming directly below Chewelah about 760 million years ago.

**HIKING QUARTZITE MOUNTAIN**

There you go. Take a hike up Quartzite Mountain and contemplate that. You are a human living on a 4.6 billion-year-old planet currently standing on the 760 million-year-old edge of ancient North America. If you are fortunate, you may live for 100 years—an infinitesimally small quantity in comparison. Don't waste it. Go for a hike. Get out there! This moderate hike is 3 miles roundtrip with 800 feet of elevation gain.

To get there from Chewelah, Wash., take Flowery Trail road east, then turn south onto Mud Lake Road. About 2.5 miles up Mud Lake Road, you will spot the Quartzite Mountain trailhead on the west (right) side of the road.

*Daniel T. Brennan is a PhD student at Curtin University in Perth, Australia. This is his first article in Out There.*



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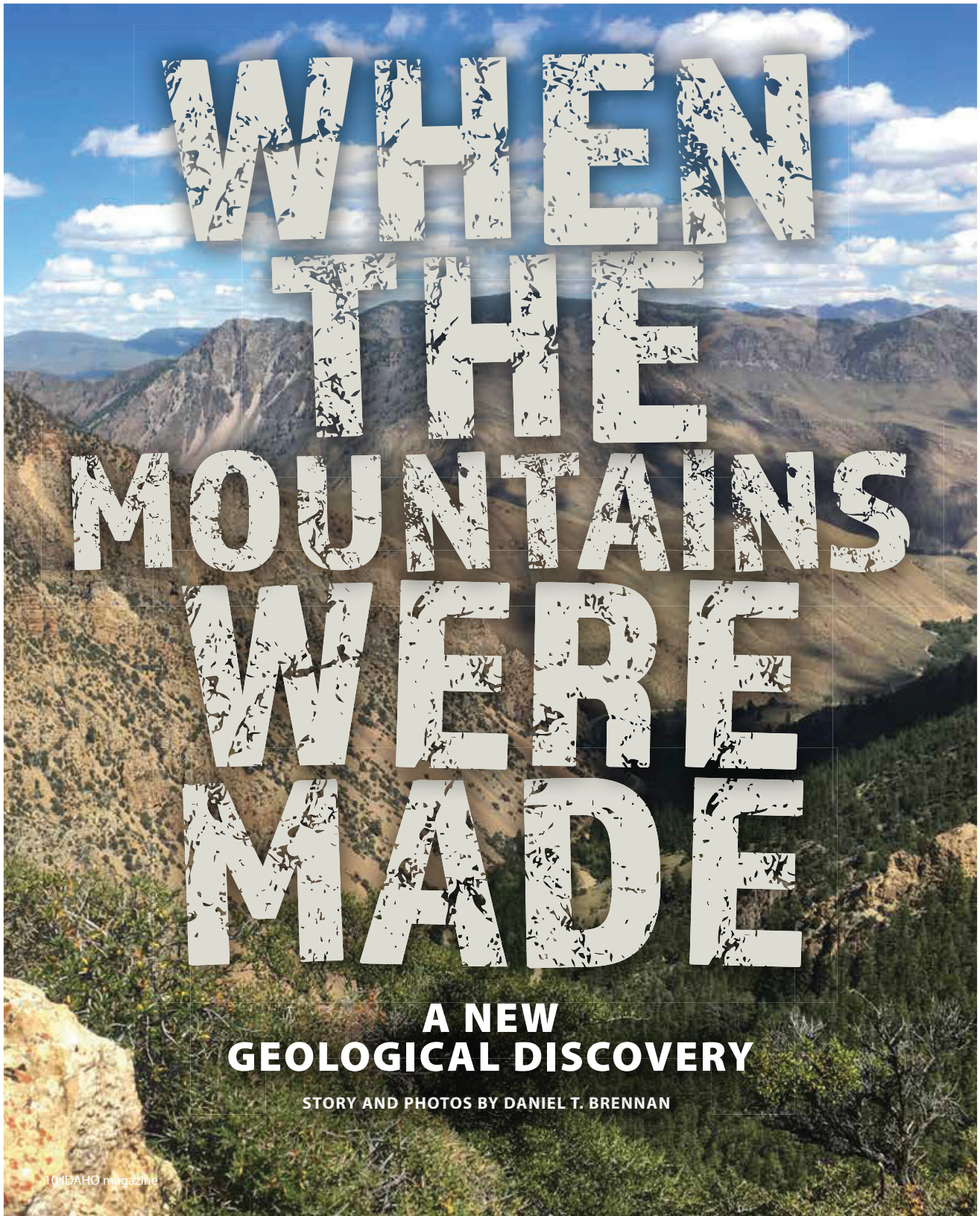
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**Appendix C2: When the Mountains Were Made: A New Geological Discovery,  
Idaho Magazine, 2020**



10 IDAHO magazine





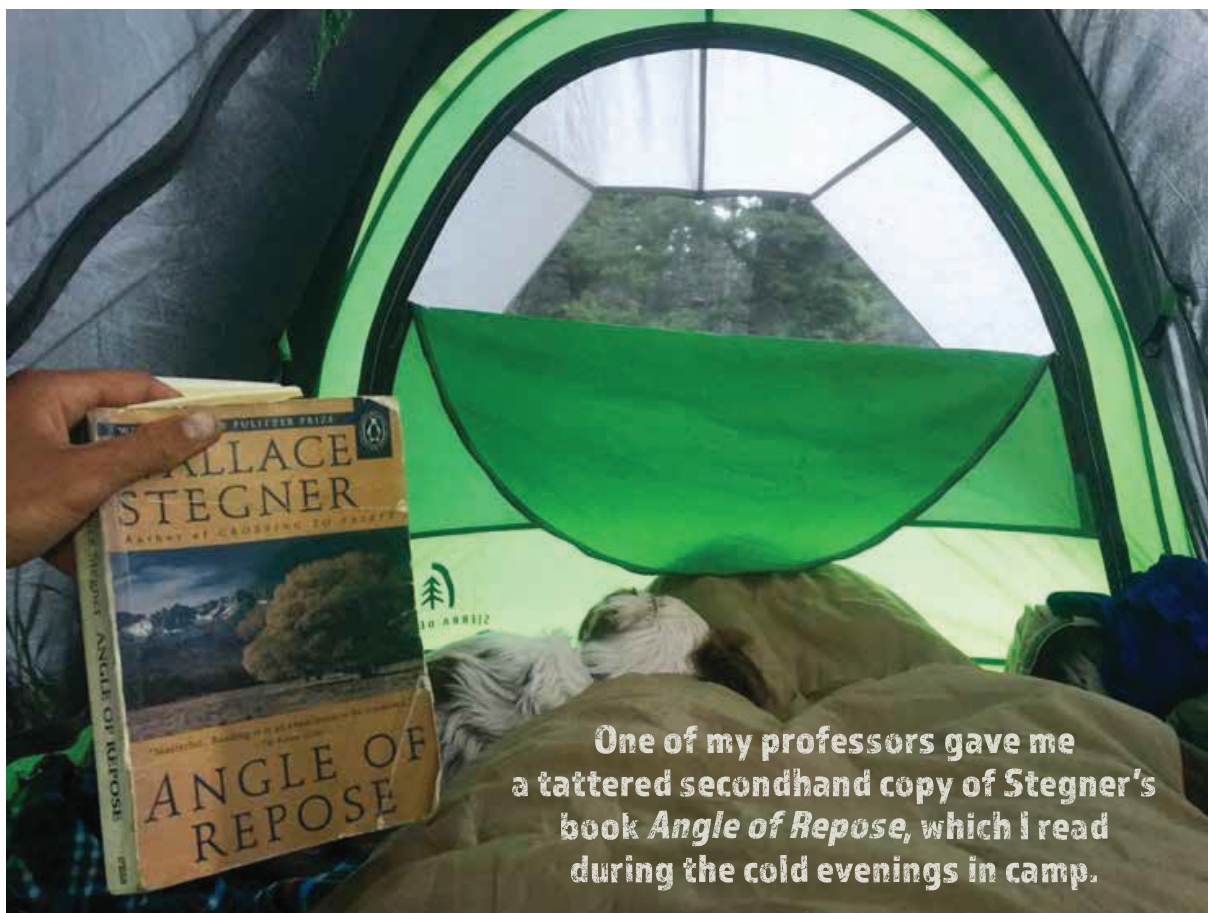
*Being footloose has always exhilarated us. It is associated in our minds with escape from history and oppression and law and irksome obligations, with absolute freedom, and the road has always led west.*

— Wallace Stegner, *The American West as Living Space*

LEFT: Dolomite cliffs near Bayhorse in the Salmon River Mountains.

ABOVE: Dan Brennan in the mountains with Blue.

On Easter of 2016, I sat in the living room with my grandfather, whom we all called Papa, when he unknowingly echoed Stegner’s phrase “the road has always led west.” I had just shared my plan to leave our home in Wisconsin and head to Idaho for graduate school, which had spurred his comment. “If you move west, you won’t move back,” Papa said nonchalantly, and quickly delved into to his own story of adventures out West on a road trip as a much younger man. He was that kind of guy, one with a story for every situation—a quality I now realize is likely the hard-earned result of a life well-lived. Between the chaos of family gatherings, Papa would share his stories to those who would listen, usually sitting in his recliner, the one with armrests worn thin from his signature way of slapping his hands when he reached a particularly exciting part, or was upset with the current politics on TV. But Papa (and Wallace Stegner) were right: the West was calling me. The day after I graduated with a bachelor’s degree, my prone-to-overheating little truck was loaded down with gear and the rising sun glinted off the rearview mirror as I headed out to start the first chapter of my own western story.



One of my professors gave me a tattered secondhand copy of Stegner's book *Angle of Repose*, which I read during the cold evenings in camp.

Actually, I had been a full year in Idaho before I was introduced to the writing of the author that Papa had unknowingly echoed. On a visit to my field area in the Salmon River Mountains, one of my professors gave me a tattered secondhand copy of Stegner's book *Angle of Repose*, which I read during the cold evenings in camp. To be honest, I spent most of that first year unsure of what I was doing . . . with both my research project and my life. I'm a geologist, and although there is no better place for young geologists to gain experience than Idaho, the billions of years of Earth's history recorded in the rocks can be daunting to a green scientist. But I knew one thing: I liked to make geologic maps, and under the tutelage of my

professors, Idaho was the place to learn. Eventually, as I had hoped, map-making became the main task of my master's research project, thanks to a bit of funding from the United States Geologic Survey (USGS). I had one summer to complete a geologic map of about a thirty-five-square-mile region in the Salmon River Mountains just west of Challis.

Short of actual hunting or gathering, I think geologic mapping is as close to the activities of our hunter-gather ancestors as a scientist can get. Maybe that's why most field geologists find something primordially satisfying about it. If you're not familiar with the premise of geologic mapping, it's relatively straightforward. Before you begin, you must lace up your hopefully

ABOVE: Dan shows off his battered copy of the novel.

broken-in boots, equip yourself with a topographic map, a rock hammer, a hand lens (which is small magnifying glass for identifying minerals), a notebook, a good compass, and as much food and water as necessary. If you have a trusty field dog, let it out of the truck. (I didn't have a dog at first, but that changed after a few weeks of isolated hikes, cold toes at night, and one trip "just to look" at dogs for adoption.) With your pack loaded—and the dog you just adopted eagerly leading the way—you start walking. The joke goes that geologic mapping is ninety percent intentional walking, but I don't think that's far from the truth.

All joking aside, geologic mapping is a scientifically rigorous activity. As I traverse any mapping area, I constantly develop, test, and alter a 3D conceptual model of the rocks beneath my feet. This geologic model is visualized in my head, sketched in my notebook, and drawn on my map.

In preparation for my Idaho field season, I poured through the literature and found I was far from the first young person who had aimed to find fame and fortune in the craggy peaks and glaciated meadows of the Salmon River Mountains. I uncovered the labor of souls kindred to mine in reading about 18th Century Shoshoni hunters who drove buffalo off the steep cliffs south of Challis, in photos of the still-standing but dilapidated late 19th-Century ruins of failed gold and silver mines at the Bayhorse ghost town, and in several late-1980s publications by a USGS geologist named Warren Hobbs. Although I was fortunate to collaborate with current Idaho Geological Survey mappers, I never got the chance to meet Warren, who died before my time. Luckily, if you're a young person driving a truck with a university logo in Challis, you open yourself up to some good-natured questioning by the locals. On a particularly warm late-summer afternoon,

one provider of such questioning happened to be an old friend of Warren's. Unfortunately, I did not catch the oldtimer's name, but with a sparkle in his eye, he made sure the university was not paying for the burger and cold drink I was scarfing down. He said Warren would have been happy to know that young geologists were still out scampering around the Salmon River Mountains, just as he had spent many a summer doing.

As I traversed the field area with my supervisors, we began to decipher something that I'm sure Warren knew all too well. In those mountains, as with most other central Idaho ranges, the original layer-cake stratigraphy of sedimentary rocks that formed in ancient lakes, rivers, and oceans was long gone. The orderly

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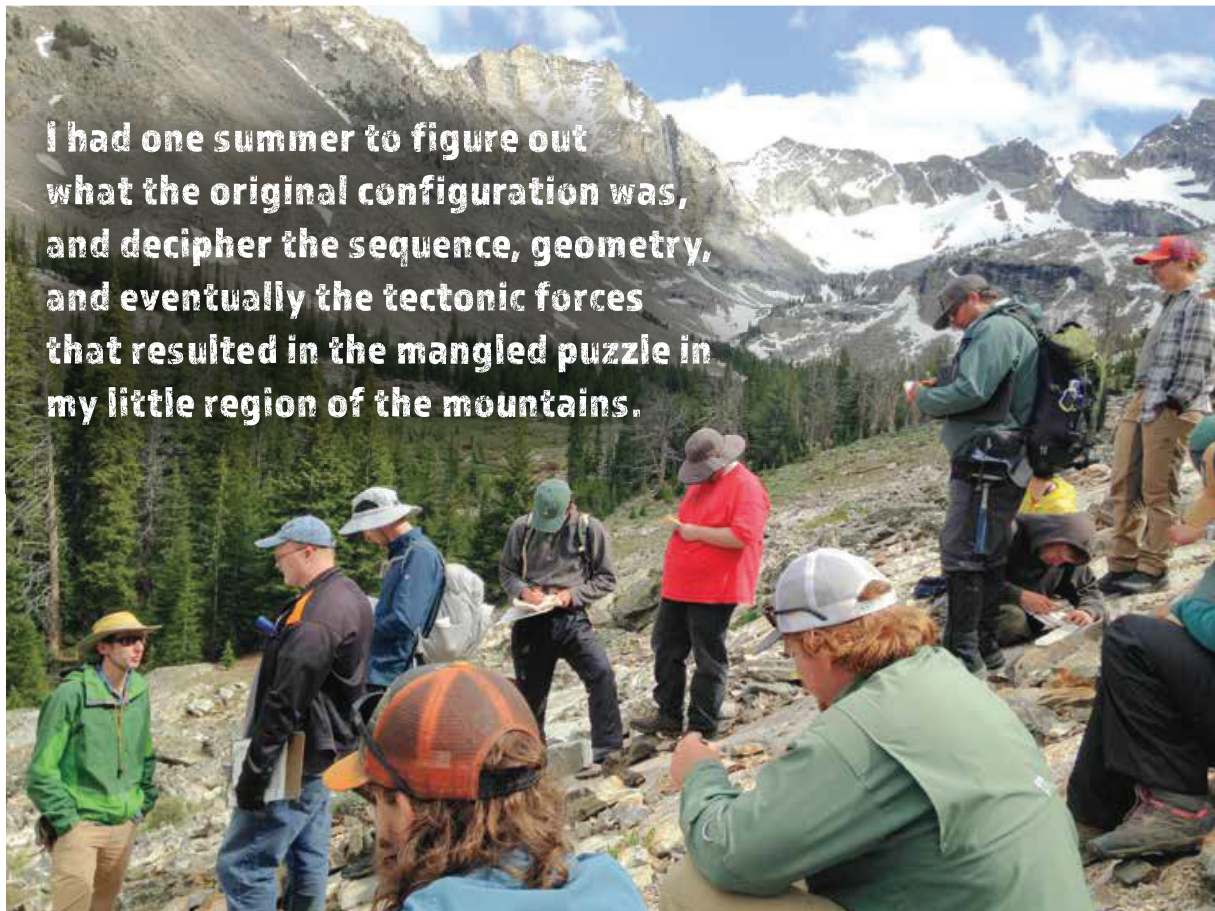
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configuration of the sedimentary rocks had been erased and deformed by younger igneous rocks, crystallized from molten material. Adjacent to where these igneous melts had intruded, the pressures and temperatures were great enough to alter the minerals in the rocks that didn't melt, a process called metamorphism. To muddy the story even more, the rocks had been displaced across a series of faults or fractures, which created a complex three-dimensional pattern. As I have learned in my travels since then, these complexities are not unique to the Salmon River Mountains, but rather are a paradox of many mountain ranges. In such cases, exposure of the ancient earth record comes at a cost: the same processes that can make the rocks available for

geologic study can mask their original identity.

I had one summer to figure out what the original configuration was, and decipher the sequence, geometry, and eventually the tectonic forces that resulted in the mangled puzzle in my little region of the mountains. I would have to piece it together one rock description at a time, and I quickly found that the secrets of ancient geologic processes entrusted to these mountains would not be shared easily. After about ten weeks of camping, hiking, eating more hot dogs than I'd like to admit, and of Blue (my newly adopted dog) terrorizing every chipmunk in Custer County, our team arrived at a hypothesis. Our geologic mapping suggested that the original geologic mapping

ABOVE: Idaho State geology class in the Pioneer Mountains.

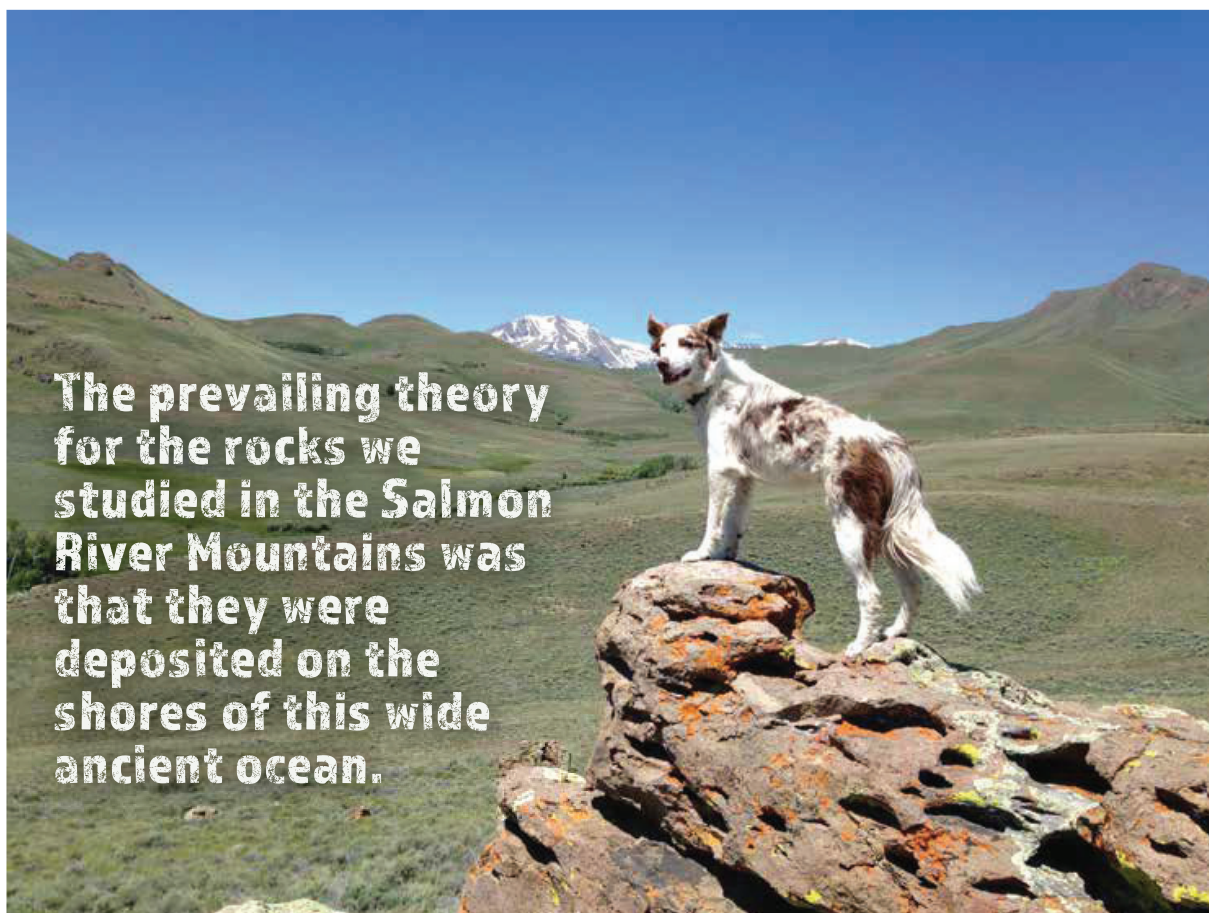
OPPOSITE, CLOCKWISE FROM TOP LEFT: Blue in the Salmon River Mountains; the author and his dog in the Pioneers; Dan in the Pioneers; Lost River Valley sunset.



done by Warren and his colleagues had incorporated a slight misunderstanding in how the rocks were configured. They had postulated a large thrust fault or fracture across which rocks were displaced. This supposed structure required that the rocks on top be older and shoved above the allegedly younger rocks they now overlaid. However, our geologic mapping suggested that instead, the rocks on top were indeed younger than those beneath, and there was very little evidence for a thrust fault.

Armed with this hypothesis, we took several kilograms of specifically chosen rock samples to the lab. We prepared the samples in our lab at Idaho State University in Pocatello but for additional analysis, we then traveled to the University of Arizona and shipped other samples off to collaborators in Wyoming, California, and





**The prevailing theory for the rocks we studied in the Salmon River Mountains was that they were deposited on the shores of this wide ancient ocean.**

Australia. This analysis consisted of measuring particular isotopes of elements such as uranium and lead in the carefully chosen minerals zircon and baddeleyite. Our results indicated that the majority of rocks thought to be about 450 million years old were actually millions of years older, and had formed mostly between 660 and 600 million years ago. As you can imagine, this was quite an exciting discovery.

The data suggested a new origin for rocks in the Salmon River Mountains. Geologists had already proposed that the shores of an ancient pre-Pacific ocean once ran through what is now Idaho. During that time, Idaho was at the edge of North America, and the majority of rocks that make up Oregon and Washington had not

yet formed. The prevailing theory for the rocks we studied in the Salmon River Mountains was that they were deposited on the shores of this wide ancient ocean. However, our data indicated that instead these rocks were deposited when the ocean was just beginning to open. Back then, other continents that were connected to the western margin of North America were in the process of slowly moving, or being rifted, away, which created an ocean. Our discovery showed that this rifting process was taking place in central Idaho more than six hundred million years ago, and this ancient ocean formed in a fashion similar to the creation of better-understood modern oceans such as the North Atlantic. Much later in

ABOVE: Hamming it up.

geological terms, a series of processes deformed the rocks of this ancient ocean shoreline into the Salmon River Mountains we see today. Warren and his colleagues had gotten in some good punches early, and had set us up for the knockout. At last, the mystery of at least some of the rocks in the Salmon River Mountains was solved.

Our uncovering of these geologic secrets would soon give me another reason to continue my westward movement. Papa seemed to have known a thing or two when he predicted that my ticket west would not be a round trip. After two years in Idaho, another geologic challenge led me across an ocean to begin a PhD program in Perth, Australia. Other geologists have suggested that Australia was one of the continents originally west of North America, whose removal might be recorded in the rocks of the Salmon River Mountains. So, I'm now doing research into that question, and although I am now on the other side of the world, geologically, it feels like home.

Unfortunately, Papa is not around anymore to let me know if my westward trajectory is destined to continue after I finish my PhD in a couple of years. But as a good mate of mine likes to say, "Life is all about stories." Wherever I go, I hope to make a few of my own and find more rocks that reveal stories of Earth's ancient past. One thing I know: at this point, if I go any farther west, I'll be headed towards home. ■

The author encourages anyone interested in studying geosciences to visit Idaho State University at [isu.edu/geosciences/undergrad](http://isu.edu/geosciences/undergrad), or get a book on geology from the local library. He recommends *Timefulness: How Thinking Like a Geologist Can Help Save the World*, by Marcia Bjornerud.

To learn more about the research Daniel and colleagues conducted, their geologic map can be downloaded at [idahogeology.org/product/t-20-01](http://idahogeology.org/product/t-20-01)

For their recent scientific paper, visit [agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2020TC006145](http://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2020TC006145)

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## **Appendix D: Supplementary Datatables**



# Appendix D2.1: SHRIMP U/Pb Results

Spot Fundamentals		SQUID OUTPUTS 27.01.2021										ISOPLOT RANGE CALCS						DTB Calcs
Fraction	Date/Time	U (ppm)	Th (ppm)	232Th	238U	4-corr	4-corr	4-corr	4-corr	207/235	206/238	207/206	Concordia Age	Discordance				
				±1σ/ma	±1σ/ma	±1σ/ma	±1σ/ma	±1σ/ma	±1σ/ma	±1σ/ma	±1σ/ma	±1σ/ma	±1σ/ma					
360B-1	2/12/2020 13:15	565.5598	343.0617	0.626853	0.205623	0.078952	5.421301	3.26006	5.226405008	0.185327	1.651189	2212.2	49.4	1724.7	27.3	2646.5	17.7	36.13
360B-10	2/12/2020 18:25	1095.304	10.57142	0.009974	0.688255	0.084385	12.08252	4.178825	4.988562229	0.113849	0.702957	2212.2	40.4	1383.1	62.1	1860.9	12.7	25.68
360B-11	2/12/2020 19:16	447.532	157.3846	0.363421	0.448084	0.099173	5.076978	3.726518	4.560969582	0.174226	0.429122	2038.2	40.3	1532.4	62.2	2597.9	17.5	41.01
360B-12	2/12/2020 19:40	648.4096	99.10715	0.157953	4.731734	1.068289	7.82059	4.095438	4.488499221	0.153573	0.813343	1847.4	38.8	1408.4	56.8	2385.3	13.9	49.06
360B-13	2/12/2020 20:05	1122.702	31.69806	0.029177	4.361709	0.053516	15.58424	6.451223	4.99077178	0.128779	1.169421	1342.4	38.2	929	43.2	2080.6	20.6	18.3
360B-14	2/12/2020 20:29	1012.067	26.49539	0.027054	0.843339	0.036368	14.80102	5.332994	4.799380555	0.112264	0.469114	1386.2	36.5	1107.9	48.9	1845.22	8.49	39.96
360B-15	2/12/2020 21:41	1554.344	14.46448	0.009617	0.564802	0.056772	24.41126	7.919583	4.523285841	0.15347	0.804088	1320.4	33.9	766.5	32.7	2384.2	13.7	67.85
360B-16	2/12/2020 22:05	1623.908	62.30459	0.039649	0.993335	0.002313	62.53068	37.02228	7.103949484	0.11658	2.07416	1366	22.7	171.8	12	1903.6	37.3	33.57
360B-17	2/12/2020 22:28	942.99	309.2797	0.338995	2.248068	0.065934	6.065685	4.665889	4.519636333	0.161676	0.346316	1780.6	38.1	1251.8	51.4	2472.5	5.85	15.79
360B-18	2/12/2020 22:51	1359.531	11.10534	0.008441	0.672572	0.02827	37.20921	6.459103	4.480721669	0.125497	0.96616	1322.3	33.9	927.9	38.7	2035	17.1	111.1
360B-19	2/12/2020 23:48	331.905	29.29865	0.091223	5.857506	0.117791	11.14903	2.977206	4.498198886	0.162512	1.074375	1275.7	41.4	1866.8	72.9	2481.2	18.1	2458.5
360B-2	3/12/2020 13:47	537.0556	159.2783	0.307351	0.496542	0.12228	4.672815	2.311236	4.587721731	0.182376	1.733992	2512.5	45.6	2317.7	89.3	2673.9	28.7	2646.1
360B-20	3/12/2020 13:42	264.5509	291.1166	0.171881	4.043984	1.13818	5.456927	2.018498	4.500491713	0.199835	0.376815	2725.3	42.7	2594.1	96.1	2824.12	6.15	13.82
360B-3	2/12/2020 14:17	2101.508	13.84426	0.006808	1.008327	0.020077	82.4959	10.84803	6.392827842	0.101411	1.114272	840.6	37.1	568.4	34.8	1649.3	20.7	57.9
360B-4	2/12/2020 15:30	2320.623	23.74162	0.005072	0.885947	0.054379	35.54229	10.25757	4.926808451	0.100554	0.864082	868.2	29.2	599.7	28.2	1634.1	16.1	63.30
360B-5	2/12/2020 15:53	982.9511	7.898468	0.008304	0.774796	0.034148	29.56924	5.10963	4.775584996	0.11279	0.41944	1418.3	36.6	1152.2	50.4	1844.03	7.59	1837.54
360B-6	2/12/2020 16:25	1724.335	16.94421	0.013741	0.522273	0.02518	52.47443	7.027994	4.602684775	0.139257	0.493973	1336.9	34.4	857.6	37	2217.21	8.56	23.9
360B-7	2/12/2020 17:14	1995.923	23.88791	0.012109	0.450077	0.035919	29.3344	8.981748	4.506307013	0.105985	0.766487	920.9	27.7	620.8	26.7	1730.6	6.1	6.74
360B-8	2/12/2020 17:37	804.4948	97.86889	0.125717	4.626751	0.133027	8.336318	3.57489	5.164520251	0.154353	1.462755	1968.6	46.7	1590	72.8	2394	24.9	2347.9
360B-9	2/12/2020 18:01	2338.375	31.00694	0.006994	0.536074	0.024691	54.15132	9.247654	4.543529281	0.105013	0.498292	56.5	28.3	661.9	28.6	1713.72	9.01	25.35
370B-10	3/12/2020 4:54	178.736	770.3403	0.673649	0.350263	0.809338	4.510497	6.373563	4.479659596	0.097498	0.3261	1562.2	35.8	1552	61.8	1579.96	6.1	1575.74
370B-11	3/12/2020 5:17	379.7817	271.6599	0.739202	0.398131	0.078464	4.56852	3.711891	4.492954361	0.098036	0.618495	1561.6	36.1	1537.8	61.5	1586.3	11.6	1584.7
370B-12	3/12/2020 5:19	515.7931	193.9921	0.388669	0.202776	0.071504	4.721745	4.015264	4.488701285	0.097515	0.573993	1492.1	35.4	1433.6	57.7	1576.3	10.3	9.63
370B-13	3/12/2020 6:02	682.8437	357.9954	0.541786	0.468089	0.079163	4.542386	3.636832	4.484133033	0.09774	0.417077	1572.2	36	1566	62.3	1580.6	7.75	1582.38
370B-14	3/12/2020 6:26	222.8164	68.77475	0.318973	1.09442	0.07897	4.970292	3.751236	4.508022518	0.098161	0.862068	1550.5	36.5	1522.7	61.1	1588.6	16.1	1584.6
370B-15	3/12/2020 6:48	1164.219	557.74	0.494629	0.338542	0.080372	4.571951	3.668401	4.480233469	0.098887	0.334844	1570.6	35.9	1554	61.9	1592.94	6.25	5.94
370B-16	3/12/2020 7:40	792.8298	370.1856	0.482515	0.162915	0.076156	5.81816	3.734862	4.583176358	0.098481	0.405957	1557	35.8	1529.4	61	1594.72	4.8	1594.72
370B-17	3/12/2020 8:04	1064.285	301.6876	0.292935	2.280893	0.069427	4.431522	4.148861	4.560780714	0.097901	0.346818	1469.7	35.5	1392	57.1	1583.68	6.49	1581.75
370B-18	3/12/2020 8:31	1839.119	67.42189	0.381208	0.364722	0.079408	4.510999	3.620998	4.478221118	0.097567	0.257153	1574.3	35.9	1572	62.5	1577.29	4.82	1577.26
370B-19	3/12/2020 9:04	1043.48	319.341	0.316259	0.26404	0.081176	5.422231	3.61991	4.483772468	0.097546	0.590159	1574.3	36.2	1572.4	62.6	1576.9	11	1576.8
370B-2	3/12/2020 9:32	2095.712	1033.705	0.509726	0.47158	0.084541	5.514265	3.472564	4.477427728	0.0977	0.240156	1609	36.2	1631.4	64.6	1579.84	4.9	1631.39
370B-20	3/12/2020 9:32	339.5627	166.8197	0.507849	1.442636	0.068738	5.269801	4.603948	4.649055605	0.097738	1.893847	1388.6	38	1267.1	53.5	1580.6	35.4	1495.4
370B-21	3/12/2020 10:43	458.8063	384.9033	0.86695	0.53999	0.076893	4.565413	4.035325	4.491758214	0.097955	0.58184	1489.7	35.4	1427.2	57.5	1579.3	10.9	9.83
370B-22	3/12/2020 11:08	2799.094	1349.131	0.498991	0.614912	0.086196	4.770723	3.520802	4.524540959	0.097811	0.205747	1598.8	36.5	1611.6	64.6	1581.96	3.85	1611.61
370B-23	3/12/2020 11:56	319.8193	298.1252	0.366301	1.08519	0.070451	5.365221	3.873448	4.501028913	0.097603	0.904434	1516.2	36.1	1480.5	59.5	1566.4	17	1560.5
370B-24	3/12/2020 12:23	1184.69	309.6723	0.270089	0.638778	0.083177	4.580189	3.801025	4.492495235	0.097324	0.317187	1587.6	36.6	1505.6	61.8	1572.03	6.26	4.26
370B-3	3/12/2020 1:00	645.5069	376.0411	0.620113	1.213404	0.076227	4.689923	3.725024	4.486586851	0.097404	0.480346	1550.4	35.9	1533	61.2	1574.16	8.99	1573.33
370B-4	3/12/2020 1:45	1359.09	124.5593	0.855104	0.304009	0.083024	4.65024	3.594009	4.493093872	0.097869	0.312623	1561.6	36.1	1582.5	63	1583.07	5.85	1583.07
370B-5	3/12/2020 2:09	1389.05	422.698	0.31447	0.406602	0.077989	4.686731	4.371237	4.637840659	0.09742	0.330692	1425.6	35.6	1328	55.7	1574.47	6.19	1572.26
370B-6	3/12/2020 2:31	1657.799	694.2526	0.43277	0.302863	0.080772	4.507358	3.672802	4.478209135	0.09759	0.476419	1563.1	35.9	1552.3	61.8	1577.73	8.92	1577.23
370B-7	3/12/2020 2:55	798.0581	51.8584	0.662807	0.832571	0.082351	4.651937	3.564775	4.482824413	0.097741	0.383781	1582.2	36.1	1594	63.3	1580.62	7.18	1593.99
370B-8	3/12/2020 3:18	366.9846	85.4325	0.240631	0.460207	0.075797	4.710276	3.875883	4.495978104	0.098128	0.606831	1524.2	35.8	1475.7	59.3	1588	11.3	1584.5
370B-9	3/12/2020 3:41	788.795	252.0717	0.330241	0.850471	0.080097	4.740558	3.64188	4.523163937	0.09816	0.416527	1574.5	36.3	1564	62.8	1588.62	7.78	1588.26
370B-1	2/12/2020 12:02	146.8367	0.234376	0.001632	6.625215	0.134285	4.240571	10.10595	5.362190765	0.062005	9.610513	75.93	80.5	58.31	3.11	173.7	3.09	91.34
460B-1	4/12/2020 2:06	1844.319	85.39201	0.047847	0.252532	0.088947	5.64358	3.400075	4.478882795	0.11252	0.247444	1742.1	37.4	1662	65.6	1839.7	4.47	1839.02
460B-10	9/12/2020 4:47	2058.108	90.72429	0.045642	0.674909	0.090546	3.361025	3.197225	2.333761334	0.113269	0.59145	1800	20.3	1764.3	35.8	1853.3	10.7	1845.76
460B-11	9/12/2020 6:17																	

# Appendix D2.2: SHRIMP U/Pb Standards

Spot Fundamentals		SQUID OUTPUTS 27.01.2021									ISOPLOT R AGE CALCS						DTB Calcs			
Fraction	Date/Time	U (ppm)	Th (ppm)	232Th /238U	4-corr 208*			4-corr 238		4-corr 207*		207/235	206/238	207/206	Concordia Age		Discordance			
					+/-1sigma	/232	+/-1sigma	/206*	+/-1sigma	/206*	1sigma Ma				1sigma Ma					
OGC-1	2/12/2020 12:49	116.8504	77.20861	0.682821	0.336222	0.185055	4.622232	1.354281	4.529620718	0.298228	0.418792	3498.2	44.7	3565	124	3460.35	6.49	3564.64	5.44	-3.02
OGC-1	8/12/2020 10:46	282.21	170.2833	0.62355	0.363085	0.190776	2.470346	1.412668	2.289737609	0.299881	0.406531	3462.1	22.8	3450.5	61.2	3468.91	6.3	3468.72	6.22	0.53
OGC-10	4/12/2020 1:47	161.1597	52.86415	0.338982	0.372119	0.188421	4.725698	1.382113	4.600009636	0.296913	0.373977	3473.9	45.3	3509	124	3453.49	5.8	3509.26	5.26	-1.61
OGC-11	4/12/2020 6:25	187.4651	154.7243	0.852922	0.278761	0.171291	4.587918	1.496572	4.537312592	0.301396	0.396712	3410.5	44.4	3299	117	3476.72	6.14	3476.27	5.74	5.11
OGC-2	2/12/2020 15:06	93.42488	60.90468	0.67369	0.889841	0.194051	4.728354	1.355634	4.539503293	0.300368	0.446708	3504.2	44.8	3562	124	3471.43	6.92	3561.91	5.93	-2.61
OGC-2	8/12/2020 11:07	257.6634	211.3093	0.847496	0.35187	0.182231	2.455159	1.407638	2.301618265	0.300919	0.427754	3469	23	3460	61.6	3474.26	6.62	3474.1	6.55	0.41
OGC-3	2/12/2020 21:17	56.1525	27.76209	0.510922	0.897618	0.187714	4.874294	1.355972	4.581727262	0.298531	0.57389	3498	45.4	3561	125	3461.92	8.89	3561.22	7.51	-2.86
OGC-3	8/12/2020 15:50	256.5092	219.6716	0.884998	0.351146	0.174733	2.477691	1.430895	2.306966564	0.298726	0.4372	3445.8	23	3416.3	61.2	3462.93	6.78	3462.38	6.63	1.35
OGC-4	3/12/2020 4:26	101.2382	52.86829	0.539663	0.390102	0.183882	4.689806	1.397894	4.540390371	0.302165	0.458757	3479.9	44.8	3479	122	3480.66	7.1	3480.65	7.08	0.05
OGC-4	8/12/2020 19:38	154.1068	134.7953	0.903909	0.428992	0.179552	2.6587	1.459878	2.387311615	0.297813	0.584039	3423.1	24.1	3363.5	62.6	3458.19	9.05	3456.33	8.68	2.74
OGC-5	3/12/2020 10:21	197.3397	155.0681	0.812044	0.253155	0.182014	4.555753	1.419381	4.507728419	0.299151	0.329236	3455.1	44.4	3438	120	3465.14	5.1	3465.09	5.05	0.78
OGC-5	8/12/2020 23:05	148.2808	84.3386	0.587778	0.473387	0.180363	2.832832	1.407987	2.379002542	0.299736	0.562848	3464.9	24	3459.3	63.7	3468.16	8.72	3468	8.61	0.26
OGC-6	3/12/2020 13:10	287.0915	245.0727	0.882158	0.2069	0.176792	4.611336	1.44555	4.536751368	0.298829	0.268484	3436.1	44.6	3389	120	3463.47	4.16	3463.38	4.04	2.15
OGC-6	9/12/2020 2:57	199.1519	92.10318	0.477927	1.224248	0.180312	3.10704	1.407321	2.342282411	0.29672	0.528268	3455.5	23.6	3460.6	62.7	3452.49	8.19	3460.61	8.01	-0.23
OGC-7	3/12/2020 15:04	171.0295	139.1057	0.840516	0.266356	0.178174	4.568067	1.444334	4.512676078	0.299662	0.703653	3439.7	44.8	3392	119	3467.8	10.9	3467.2	10.6	2.19
OGC-7	9/12/2020 7:01	88.31484	45.40984	0.531359	0.622478	0.191301	3.21373	1.416842	2.513854176	0.304488	0.71615	3474.2	25.7	3442.6	67.1	3492.5	11.1	3491.2	10.8	1.43
OGC-8	3/12/2020 18:51	161.6706	84.26102	0.538601	0.31652	0.183565	4.61469	1.376344	4.516208737	0.300984	0.375268	3491.3	44.5	3521	123	3474.6	5.81	3520.6	5.36	-1.34
OGC-9	3/12/2020 23:36	90.48369	60.67674	0.692985	0.719676	0.186127	4.736919	1.394378	4.560978392	0.297796	0.538906	3468.1	45.1	3485	123	3458.1	8.35	3485.42	7.94	-0.78

# Appendix D2.3: SIMS O- Results

Analysis session 1		20 µm diameter, 2.5 nA, Cs <sup>+</sup> ion-beam						
Sample #	Analysis #	Raw Data from CIPS		SIMS corrected		Drift corrected		Notes
		18O/16O	1σ error	d18O	± per mil	OH/O	± rel (%)	
37DTB19	OH-DB_09-D820_37DTB19@01	0.002016266	3.06086E-07	5.51884379	0.151808103	0.000857162	0.3508533	
37DTB19	OH-DB_09-37DB19@02	0.002016441	3.32346E-07	5.606012346	0.164818147	0.000759828	0.1151747	
37DTB19	OH-DB_09-37DB19@03	0.002015946	2.81726E-07	5.359034772	0.139748723	0.000745774	0.1942408	
37DTB19	OH-DB_09-37DB19@04	0.00201548	2.88722E-07	5.126585291	0.143252387	0.000795772	0.4011484	
37DTB19	OH-DB_09-37DB19@05	0.002015897	3.39436E-07	5.334487305	0.168379672	0.00076661	0.2437459	
37DTB19	OH-DB_09-37DB19@06	0.002015423	2.99935E-07	5.098030075	0.148819812	0.000986084	0.2392039	
37DTB19	OH-DB_09-37DB19@07	0.002015292	2.84799E-07	5.032904143	0.141319191	0.000808253	0.567568	
37DTB19	OH-DB_09-37DB19@08	0.002016017	3.16431E-07	5.394603551	0.156958348	0.000861948	0.5737716	
37DTB19	OH-DB_09-37DB19@09	0.002015448	3.18059E-07	5.110554292	0.15781074	0.000858881	0.4914057	
37DTB19	OH-DB_09-37DB19@10	0.002014307	2.91251E-07	4.541453839	0.144591377	0.000766662	0.1029989	
37DTB19	OH-DB_09-37DB19@11	0.002014643	2.69003E-07	4.709278356	0.133524047	0.000820311	0.5304983	
37DTB19	OH-DB_09-37DB19@12	0.00201529	3.03855E-07	5.031902205	0.150774969	0.000748316	0.1278924	
37DTB19	OH-DB_09-37DB19@13	0.002015458	2.85052E-07	5.11556398	0.14143272	0.000792465	0.2259297	
37DTB19	OH-DB_09-37DB19@14	0.002015777	3.22489E-07	5.274872029	0.159982353	0.000753895	0.4566869	
37DTB19	OH-DB_09-37DB19@15	0.002016066	3.42332E-07	5.419151017	0.169801831	0.000716048	0.2629586	
37DTB19	OH-DB_09-37DB19@16	0.002016213	3.21781E-07	5.492292449	0.139458705	0.000804882	0.363914	
37DTB19	OH-DB_09-37DB19@17	0.002015205	3.00356E-07	4.989319865	0.149044993	0.000792993	0.2558843	
37DTB19	OH-DB_09-37DB19@18	0.002015534	3.08184E-07	5.153637601	0.139837888	0.000799125	0.1219552	
37DTB19	OH-DB_09-37DB19@19	0.002014837	3.21374E-07	4.805965317	0.159503905	0.000902445	0.2574412	
37DTB19	OH-DB_09-37DB19@20	0.002015767	4.09255E-07	5.269862342	0.203027147	0.000745844	0.1927767	
46DTB19	OH-DB_09-46DB19@01	0.002015749	3.12402E-07	5.260844905	0.154980417	0.000791339	0.4824054	
46DTB19	OH-DB_09-46DB19@02	0.002016283	3.83631E-07	5.527360259	0.190266324	0.000756203	0.3999153	
46DTB19	OH-DB_09-46DB19@03	0.002016438	2.90515E-07	5.60450944	0.144073574	0.000774367	0.5494097	
46DTB19	OH-DB_09-46DB19@04	0.002016929	3.08165E-07	5.849483139	0.152788979	0.000770785	0.197718	
46DTB19	OH-DB_09-46DB19@05	0.002017005	3.17804E-07	5.887055792	0.157562422	0.000821509	0.5677005	
46DTB19	OH-DB_09-46DB19@06	0.002016197	2.98799E-07	5.48427695	0.148199406	0.000786919	0.2243366	
46DTB19	OH-DB_09-46DB19@07	0.002015497	3.05506E-07	5.135101759	0.151578558	0.000867206	0.4198946	
46DTB19	OH-DB_09-46DB19@08	0.002015671	2.90978E-07	5.221769346	0.144358029	0.001003019	0.9196111	
46DTB19	OH-DB_09-46DB19@09	0.002017	2.71523E-07	5.884550948	0.144617258	0.001395404	4.015526	omitted high OH/O
46DTB19	OH-DB_09-46DB19@10	0.002016567	3.27575E-07	5.686833435	0.162441742	0.000872832	0.1851135	
46DTB19	OH-DB_09-46DB19@11	0.002016368	2.9277E-07	5.569441653	0.145196642	0.000921191	0.5056991	
46DTB19	OH-DB_09-46DB19@12	0.002017166	2.95829E-07	5.967711354	0.144665894	0.002378079	2.745597	omitted high OH/O
46DTB19	OH-DB_09-46DB19@13	0.002016505	2.66461E-07	5.638074343	0.132140071	0.001005206	0.217211	
46DTB19	OH-DB_09-46DB19@14	0.002016386	2.58765E-07	5.578459067	0.128331081	0.001186586	0.6212962	omitted high OH/O
46DTB19	OH-DB_09-46DB19@15	0.002015347	3.72267E-07	5.060457422	0.184716174	0.000943951	0.7896001	
46DTB19	OH-DB_09-46DB19@16	0.002015323	3.02353E-07	5.048434173	0.150523469	0.001043421	0.4183532	
46DTB19	OH-DB_09-46DB19@17	0.002015864	2.98742E-07	5.317955338	0.148195591	0.000856083	0.2628726	
46DTB19	OH-DB_09-46DB19@18	0.002015603	3.30695E-07	5.188204442	0.164067722	0.000910641	0.2741711	
46DTB19	OH-DB_09-46DB19@19	0.002016272	2.71938E-07	5.521849603	0.134871544	0.000886009	0.1250535	
46DTB19	OH-DB_09-46DB19@20	0.002016962	3.23785E-07	5.865514137	0.16053106	0.000882343	0.2237801	
67DTB19	OH-DB_09-67DB19@01	0.002020957	2.92151E-07	7.887866694	0.194042022	0.001780457	4.875086	omitted high OH/O
67DTB19	OH-DB_09-67DB19@02	0.002016982	2.74882E-07	5.875533511	0.136283737	0.00071295	0.5009309	
67DTB19	OH-DB_09-67DB19@03	0.002018159	2.84159E-07	6.426268829	0.14080088	0.001604421	4.486829	omitted high OH/O
67DTB19	OH-DB_09-67DB19@04	0.002016811	2.87526E-07	5.790368831	0.142564532	0.000914548	1.120014	
67DTB19	OH-DB_09-67DB19@05	0.002012857	2.90348E-07	3.818555991	0.144246603	0.000710549	5.998934	
67DTB19	OH-DB_09-67DB19@06	0.002016481	3.16902E-07	5.626051094	0.157155752	0.000679686	0.1958074	
67DTB19	OH-DB_09-67DB19@07	0.002016347	2.98094E-07	5.558921287	0.14783864	0.000711895	0.4390393	
67DTB19	OH-DB_09-67DB19@08	0.002016024	3.17266E-07	5.398110332	0.157372094	0.000706171	3.334309	
67DTB19	OH-DB_09-67DB19@09	0.002016131	3.8142E-07	5.451213015	0.189184371	0.000808887	1.802832	
67DTB19	OH-DB_09-67DB19@10	0.002015924	2.80113E-07	5.348013461	0.138950289	0.000725139	0.3149293	
67DTB19	OH-DB_09-67DB19@11	0.002015411	3.24853E-07	5.09201845	0.161184488	0.000703479	0.218183	
67DTB19	OH-DB_09-67DB19@12	0.002016514	3.24684E-07	5.642082093	0.161012678	0.000762307	0.2272872	
67DTB19	OH-DB_09-67DB19@13	0.002016064	3.34509E-07	5.41814908	0.165921817	0.000672457	0.3035271	
67DTB19	OH-DB_09-67DB19@14	0.002016162	2.81184E-07	5.466743045	0.139465173	0.00069224	0.2536313	
67DTB19	OH-DB_09-67DB19@15	0.0020153	3.536E-07	5.036911892	0.175457584	0.000701259	0.2124638	
67DTB19	OH-DB_09-67DB19@16	0.002015918	2.75734E-07	5.345007648	0.136778464	0.000690229	0.3118861	
67DTB19	OH-DB_09-67DB19@17	0.002012996	3.23452E-07	3.887689673	0.160681932	0.000759254	0.7287932	
67DTB19	OH-DB_09-67DB19@18	0.002015898	3.4336E-07	5.334988274	0.170326268	0.000774998	1.460556	
67DTB19	OH-DB_09-67DB19@19	0.0020153	2.66571E-07	5.036911892	0.132273386	0.000783027	0.1887783	
67DTB19	OH-DB_09-67DB19@20	0.002015229	3.4611E-07	5.001343114	0.171747071	0.000790339	0.5659688	

Analysis session 2		10 µm diameter, 1.0 nA, Cs <sup>+</sup> ion-beam						
Sample #	Analysis #	Raw Data from CIPS		SIMS corrected		Drift corrected		Notes
		18O/16O	1σ error	d18O	± per mil	OH/O	± rel (%)	
52DTB19	OH-DB_52DTB19@1	0.00201456	3.65128E-07	4.667697953	0.181244599	0.000951835	0.187781	
52DTB19	OH-DB_52DTB19@2	0.002013571	3.23901E-07	4.174744744	0.160859222	0.001172443	0.5562904	
52DTB19	OH-DB_52DTB19@3	0.002014413	3.05061E-07	4.594556522	0.151439222	0.000997259	0.2089095	
52DTB19	OH-DB_52DTB19@4	0.002014449	3.20391E-07	4.633131113	0.15904311	0.000952211	0.2070688	
52DTB19	OH-DB_52DTB19@5	0.002012897	2.97348E-07	3.83859474	0.147721254	0.000975512	0.1273858	
52DTB19	OH-DB_52DTB19@6	0.002014484	3.00463E-07	4.6301253	0.14915154	0.000943785	0.1730374	
52DTB19	OH-DB_52DTB19@7	0.00201525	2.9129E-07	5.011863457	0.144542949	0.001173869	0.6260424	
52DTB19	OH-DB_52DTB19@8	0.002014478	3.40544E-07	4.62719488	0.169048024	0.001013108	0.2638681	

# Appendix D2.4: SIMS O- Standards

Analysis session 1		20 μm diameter, 2.5 nA, Cs <sup>+</sup> ion-beam						
Sample #	Analysis #	Raw Data from CIPS		SIMS corrected		Drift corrected		Notes
		18O/16O	1sigma error	d18O	± per mil	OH/O	± rel (%)	
91500	OH-DB_09-DB20_91500@4	0.002025401	2.73962E-07	10.07415226	0.135263313	0.000687249	0.5323319	
91500	OH-DB_09-DB20_91500@3	0.002025391	3.3901E-07	10.06914258	0.16738018	0.000656006	0.2116581	
91500	OH-DB_09-DB20_91500@2	0.002025256	2.84147E-07	10.00401664	0.140301653	0.000661941	0.4074392	
91500	OH-DB_09-DB20_91500@14	0.002025039	2.95929E-07	9.893803528	0.14613504	0.000670406	0.1463796	
91500	OH-DB_09-DB20_91500@13	0.002025428	3.43505E-07	10.08767842	0.169596216	0.000652035	0.1458913	
91500	OH-DB_09-DB20_91500@12	0.002024792	3.19069E-07	9.770565226	0.157580972	0.000670077	0.3364866	
91500	OH-DB_09-DB20_91500@11	0.002025348	2.71902E-07	10.04810189	0.134249664	0.000676422	0.5313374	
91500	OH-DB_09-DB20_91500@09	0.002025177	3.48236E-07	9.962436241	0.171953366	0.000692961	0.5174865	
91500	OH-DB_09-DB20_91500@08	0.002024948	2.77273E-07	9.848215376	0.136928478	0.000662578	0.1474669	
91500	OH-DB_09-DB20_91500@07	0.002025252	2.81412E-07	9.973958522	0.138955236	0.000677362	0.1864196	
91500	OH-DB_09-DB20_91500@06	0.002025184	2.94617E-07	9.965943022	0.145476677	0.000664564	0.3690465	
91500	OH-DB_09-DB20_91500@05	0.002024706	2.79989E-07	9.727481917	0.138286387	0.000694405	0.5311041	
91500	OH-DB_09-DB20_91500@010	0.002025113	3.50732E-07	9.930875213	0.173191217	0.000658301	0.1500499	
91500	OH-DB_09-DB20_91500@01	0.00204858	3.06937E-07	9.803629161	0.151584245	0.000665258	0.3946493	
OGC	OH-DB_09-DB20_OGC@13	0.002016687	3.08647E-07	5.72874968	0.153046334	0.001089232	0.140743	
OGC	OH-DB_09-DB20_OGC@12	0.002016559	3.1645E-07	5.664625685	0.156925875	0.001043832	0.1129922	
OGC	OH-DB_09-DB20_OGC@11	0.00201674	3.98285E-07	5.754800053	0.197489386	0.000948023	0.3083848	
OGC	OH-DB_09-DB20_OGC@10	0.002016268	2.86055E-07	5.519845728	0.141873554	0.000981907	0.1130113	
OGC	OH-DB_09-DB20_OGC@09	0.002016035	3.18542E-07	5.403620987	0.158003934	0.000957154	0.2112503	
OGC	OH-DB_09-DB20_OGC@08	0.002015189	2.91543E-07	4.981304365	0.144672793	0.000915684	0.1470667	
OGC	OH-DB_09-DB20_OGC@07	0.002016805	3.18388E-07	5.787363019	0.15788676	0.00100767	0.209695	
OGC	OH-DB_09-DB20_OGC@06	0.002016358	2.90876E-07	5.564431943	0.144257908	0.000893493	0.1580112	
OGC	OH-DB_09-DB20_OGC@05	0.002016569	3.54759E-07	5.669635372	0.175921933	0.00083367	0.2175937	
OGC	OH-DB_09-DB20_OGC@04	0.002016356	3.43124E-07	5.563430006	0.170170348	0.000856599	0.2124317	
OGC	OH-DB_09-DB20_OGC@03	0.0020179	2.6318E-07	6.33418912	0.13042275	0.000925105	0.3843214	
OGC	OH-DB_09-DB20_OGC@02	0.002016758	3.14937E-07	5.763817489	0.156160053	0.000819279	0.4766153	
OGC	OH-DB_09-DB20_OGC@01	0.002017293	3.10541E-07	6.030833811	0.153939654	0.000986527	0.225265	
M257	OH-DB_09-DB20_M257@4	0.002033391	2.98287E-07	14.05885738	0.146694184	0.000643488	0.2676096	
M257	OH-DB_09-DB20_M257@30	0.002033596	3.25667E-07	14.16105499	0.16014357	0.000654028	0.195474	
M257	OH-DB_09-DB20_M257@3	0.002033235	2.78309E-07	13.98120723	0.136879808	0.000650297	0.1625557	
M257	OH-DB_09-DB20_M257@29	0.002033071	2.99631E-07	13.89954933	0.147378253	0.000662105	0.1113873	
M257	OH-DB_09-DB20_M257@28	0.002032616	3.08013E-07	13.6726105	0.151535136	0.0006703	0.3621382	
M257	OH-DB_09-DB20_M257@27	0.002033152	3.02482E-07	13.93962682	0.148774856	0.000638871	0.1699941	
M257	OH-DB_09-DB20_M257@26	0.002032675	3.33561E-07	13.70166669	0.164099626	0.000670975	0.398922	
M257	OH-DB_09-DB20_M257@25	0.002033107	3.07945E-07	13.91708323	0.151465412	0.000641157	0.1556723	
M257	OH-DB_09-DB20_M257@24	0.002032869	2.88182E-07	13.79885462	0.141761172	0.000642864	0.3252287	
M257	OH-DB_09-DB20_M257@23	0.002033209	3.02998E-07	13.96818204	0.149024494	0.000646363	0.2459461	
M257	OH-DB_09-DB20_M257@22	0.002033046	3.24747E-07	13.88702511	0.15973423	0.00067389	0.3661834	
M257	OH-DB_09-DB20_M257@21	0.002032833	3.28822E-07	13.78081974	0.161755516	0.000675872	0.255695	
M257	OH-DB_09-DB20_M257@20	0.002033173	3.3958E-07	13.95014717	0.167019848	0.000658617	0.2600312	
M257	OH-DB_09-DB20_M257@2	0.00203327	3.98378E-07	13.99874113	0.14674806	0.000648604	0.1942788	
M257	OH-DB_09-DB20_M257@19	0.002033451	3.32724E-07	14.0889155	0.163625035	0.000651835	0.1603915	
M257	OH-DB_09-DB20_M257@18	0.00203246	2.72836E-07	13.59445938	0.134239145	0.000680224	0.2829176	
M257	OH-DB_09-DB20_M257@17	0.00203294	2.60471E-07	13.83392143	0.128125275	0.000667519	0.3641615	
M257	OH-DB_09-DB20_M257@16	0.002033281	2.77544E-07	14.00425179	0.136500688	0.000658583	0.2865007	
M257	OH-DB_09-DB20_M257@15	0.002033245	3.22523E-07	13.98621691	0.158624509	0.000642591	0.1161775	
M257	OH-DB_09-DB20_M257@14	0.002033524	3.00555E-07	14.12548622	0.147800214	0.000645779	0.2443702	
M257	OH-DB_09-DB20_M257@13	0.002033435	3.36534E-07	14.08841453	0.165498782	0.000650669	0.1970071	
M257	OH-DB_09-DB20_M257@12	0.002034067	3.23027E-07	14.39600932	0.158808488	0.000647806	0.2153644	
M257	OH-DB_09-DB20_M257@11	0.002033583	3.03471E-07	14.1545424	0.1492296	0.000648651	0.2568071	
M257	OH-DB_09-DB20_M257@10	0.002033712	3.83584E-07	14.21916736	0.188612903	0.000667922	0.2456096	
M257	OH-DB_09-DB20_M257@09	0.002033519	3.1474E-07	14.12298137	0.154776226	0.000659663	0.2778733	
M257	OH-DB_09-DB20_M257@08	0.002033513	2.77601E-07	14.11997556	0.136512877	0.000667268	0.2093164	
M257	OH-DB_09-DB20_M257@07	0.002033864	3.65642E-07	14.29481364	0.179777259	0.00064654	0.2257004	
M257	OH-DB_09-DB20_M257@06	0.002033357	3.57767E-07	14.04182444	0.175948953	0.000660983	0.2102762	
M257	OH-DB_09-DB20_M257@05	0.002033608	3.22347E-07	14.16706662	0.158509873	0.000624511	0.41315	
M257	OH-DB_09-DB20_M257@02	0.002033185	3.24303E-07	13.95615879	0.159504744	0.000643087	0.231571	
M257	OH-DB_09-DB20_M257@01	0.002033329	3.22878E-07	14.02779732	0.15879291	0.000660593	0.3742145	
M257	OH-DB_09-DB20_M257@01	0.002033259	3.25413E-07	13.99323048	0.160045095	0.000655022	0.2275834	

Analysis session 2		10 μm diameter, 1.0 nA, Cs <sup>+</sup> ion-beam						
Sample #	Analysis #	Raw Data from CIPS		SIMS corrected		Drift corrected		Notes
		18O/16O	1sigma error	d18O	± per mil	OH/O	± rel (%)	
M257	OH-DB_52DTB19_M257@7	0.002033513	3.46641E-07	14.11997556	0.170464127	0.0008626	0.2054075	
M258	OH-DB_52DTB19_M257@6	0.002033424	3.18243E-07	14.07538934	0.156505899	0.000900419	0.1926557	
M259	OH-DB_52DTB19_M257@5	0.00203344	3.04378E-07	14.08340484	0.149686074	0.000975826	0.4281997	
M260	OH-DB_52DTB19_M257@4	0.002033133	2.98886E-07	13.93010842	0.147007435	0.000949523	0.3940813	
M261	OH-DB_52DTB19_M257@2	0.002033196	2.88108E-07	13.96166945	0.141702169	0.000946904	0.1929651	
M262	OH-DB_52DTB19_M257@01	0.002032816	3.45779E-07	13.77230328	0.170098706	0.000957122	0.1667874	
91500	OH-DB_52DTB19_91500@01	0.002025602	3.24555E-07	10.17434601	0.160226595	0.001010495	0.4473895	omitted sample topology
91500	OH-DB_52DTB19_91500@2	0.002025449	3.25421E-07	10.09819876	0.165602069	0.001012409	0.5280014	omitted sample topology
91500	OH-DB_52DTB19_91500@3	0.002025547	3.19891E-07	10.1472937	0.157928269	0.000988545	0.3153125	omitted sample topology
91500	OH-DB_52DTB19_91500@4	0.002025569	3.8907E-07	10.15831501	0.142710472	0.000825623	0.1556147	omitted sample topology
91500	OH-DB_52DTB19_91500@5	0.002025503	2.9946E-07	10.12525107	0.147844556	0.000823978	0.3679922	omitted sample topology
OGC	OH-DB_52DTB19_OGC@01	0.002017058	2.9771E-07	5.913607133	0.147596137	0.001304474	0.2429095	
OGC	OH-DB_52DTB19_OGC@03	0.002017023	3.38173E-07	5.896073229	0.167659562	0.001092478	0.1834291	
OGC	OH-DB_52DTB19_OGC@2	0.002016708	3.38891E-07	5.739270023	0.168041624	0.001308865	0.5039747	











# Appendix D2.7: LA-ICPMS Standards

Curtin University Geochronology center	Sample ID	Analysis date-time	Spot ID	238U conc.				235U conc.				LA-ICP-MS Isotopic Ratios					Apparent Ages - Calculated in Isoplot R (Ma)				Discard
				ppm		ppm		238U/235U	±2σ P/CP	207Pb/235U	±2σ P/CP	rMs	238U/235U	±2σ P/CP	207Pb/235U	±2σ P/CP					
				ppm	ppm	ppm	ppm														
Standard Results - U/Pb on 8900																					
Z_91500_1	17/03/2021 (4)	91500-1	80.2	2.2	30.02	0.83000	5.58975	0.04899386	0.0744	0.0022	0.2058	1560.95	8.75	1051.4	59.6	-0.91					
Z_91500_2	17/03/2021 (4)	91500-2	80.5	2.2	30.13	0.83000	5.56709	0.04902093	0.075	0.0023	0.2127	1564.79	8.74	1057.6	61.6	0.26					
Z_91500_3	17/03/2021 (4)	91500-3	80.3	2.4	30.38	0.83000	5.56709	0.04902093	0.0746	0.0023	0.2094	1562.06	8.75	1056.8	62.1	-0.50					
Z_91500_4	17/03/2021 (4)	91500-4	79.8	2.2	29.58	0.83000	5.56902	0.04935602	0.0742	0.0023	0.2025	1561.51	8.75	1099.4	60.4	1.45					
Z_91500_5	17/03/2021 (4)	91500-5	79.4	2.2	29.56	0.83000	5.57745	0.04979006	0.0748	0.0022	0.2062	1563.15	8.75	1062.2	59.2	-0.09					
Z_91500_6	17/03/2021 (4)	91500-6	80.6	2.2	30.31	0.83000	5.58755	0.04946197	0.0752	0.0022	0.2044	1560.96	8.2	1072.9	58.8	1.11					
Z_91500_7	17/03/2021 (4)	91500-7	79.4	2.2	29.77	0.83000	5.56735	0.04940404	0.0743	0.0022	0.2123	1565.88	8.74	1048.7	59.7	-1.64					
Z_91500_8	17/03/2021 (4)	91500-8	80.6	2.1	30.31	0.83000	5.59391	0.0491981	0.0745	0.0024	0.2185	1560.42	8.2	1054.1	64.0	-0.60					
Z_91500_9	17/03/2021 (4)	91500-9	81.6	2.2	30.66	0.83000	5.56683	0.04954774	0.0744	0.0022	0.2078	1565.34	8.74	1051.4	59.6	-1.33					
Z_91500_10	17/03/2021 (4)	91500-10	78	2.3	29.78	0.83000	5.59904	0.05015995	0.0754	0.0023	0.1789	1559.32	8.75	1078.3	61.2	-1.76					
Z_91500_11	17/03/2021 (4)	91500-11	80.2	2.1	30.23	0.75000	5.58357	0.04984642	0.076	0.0024	0.2406	1562.6	8.75	1094.2	61.2	2.89					
Z_91500_12	17/03/2021 (4)	91500-12	79.5	2.1	29.78	0.75000	5.57416	0.04911319	0.0738	0.0022	0.11229	1563.7	8.75	1051.1	60.2	-2.76					
Z_91500_13	17/03/2021 (4)	91500-13	80.2	2.1	30.23	0.75000	5.58357	0.04984642	0.076	0.0024	0.2406	1562.6	8.75	1094.2	61.2	2.89					
Z_91500_14	18/03/2021 (5)	91500-14	80.9	2.1	30.36	0.78000	5.58592	0.04861502	0.0745	0.0023	0.2364	1561.51	8.2	1096.0	60.5	1.22					
Z_91500_15	18/03/2021 (5)	91500-17	79.4	2	29.76	0.75000	5.57745	0.05287962	0.0751	0.0023	0.30092	1563.15	9.29	1051.8	61.7	-4.66					
Z_O61_1	17/03/2021 (4)	061-1	174.4	5.1	137.2	5.30000	1.42959	0.06339932	0.299	0.0014	0.37891	3419	121	3464.35	7.26	3.31					
Z_O61_2	17/03/2021 (4)	061-2	201	12	293	17.00000	1.39879	0.06362126	0.295	0.0014	0.44315	3471	120	3465.39	7.25	-0.34					
Z_O61_3	17/03/2021 (4)	061-3	195	5.4	139	21.00000	1.47322	0.06739935	0.292	0.0014	0.31639	3411	119	3466.39	7.25	1.59					
Z_O61_4	17/03/2021 (4)	061-4	188.9	4.3	86.4	3.20000	1.4041	0.0630879	0.287	0.0015	0.2752	3467	120	3462.8	7.78	-0.12					
Z_O61_5	17/03/2021 (4)	061-5	188.9	2.9	146.5	1.70000	1.43494	0.06566209	0.292	0.0016	0.39773	3413	121	3465.39	8.29	1.51					
Z_O61_6	17/03/2021 (4)	061-6	188.7	6.5	234.9	8.30000	1.38979	0.06302558	0.289	0.0015	0.27571	3510	123	3463.84	7.78	-0.33					
Z_O61_7	17/03/2021 (4)	061-7	188.6	3.5	119.7	4.20000	1.46605	0.06662195	0.291	0.0017	0.40007	3352	119	3465.39	8.12	1.27					
Z_O61_8	17/03/2021 (4)	061-8	142.1	4.5	141.3	4.20000	1.41825	0.06535361	0.297	0.0016	0.36317	3416	121	3467.98	8.27	1.50					
Z_O61_9	17/03/2021 (4)	061-9	115.9	5.1	112.3	5.50000	1.42744	0.06429189	0.295	0.0015	0.35312	3441	122	3462.36	7.79	0.60					
Z_O61_10	17/03/2021 (4)	061-10	146.2	4	105.3	2.90000	1.39278	0.06401254	0.294	0.0017	0.35595	3489	124	3466.43	8.8	-0.05					
Z_O61_11	17/03/2021 (4)	061-11	205.5	4.1	125.5	5.20000	1.44669	0.0669998	0.288	0.0014	0.296	3387	122	3463.2	7.81	3.20					
Z_O61_12	17/03/2021 (4)	061-12	144	3.4	145.7	4.00000	1.40721	0.06337131	0.294	0.0016	0.47642	3461	122	3466.43	8.28	1.20					
Z_O61_13	18/03/2021 (5)	061-13	163	3.7	164.9	3.90000	1.46531	0.0665303	0.289	0.0017	0.30247	3353	119	3463.84	7.86	2.20					
Z_O61_14	18/03/2021 (5)	061-17	124.3	6.4	165.3	8.20000	1.46579	0.06332333	0.295	0.0015	0.32313	3464	121	3464.07	7.77	0.23					
Z_G11_1	17/03/2021 (4)	011-1	369	11	91.8	0.27000	10.7475	0.07783009	0.06219	0.0016	0.23203	603.62	4.4	609.4	57.5	0.95					
Z_G11_2	17/03/2021 (4)	011-2	372	11	91	0.27000	10.11941	0.07884888	0.06991	0.0015	0.15607	607.49	4.52	599.4	54.2	1.35					
Z_G11_3	17/03/2021 (4)	011-3	367	10	90.2	0.26000	10.1216	0.07790517	0.06936	0.0016	0.129	603.98	4.4	615.5	57.2	0.97					
Z_G11_4	17/03/2021 (4)	011-4	365	11	85.4	0.27000	10.15495	0.07884817	0.06968	0.0016	0.24131	605.49	4.46	623.4	57	2.87					
Z_G11_5	17/03/2021 (4)	011-5	368	9.9	90.8	0.34000	10.14399	0.07811334	0.06622	0.0016	0.09888	606.2	4.34	610.5	57.4	1.30					
Z_G11_6	17/03/2021 (4)	011-6	314	9.2	91.4	0.27000	10.12047	0.07803416	0.06977	0.0016	0.201	602.79	4.46	594.3	58	-1.43					
Z_G11_7	17/03/2021 (4)	011-7	369	11	83.4	0.26000	10.32326	0.07674835	0.06978	0.0016	0.21895	599.92	4.29	626.9	56.8	4.30					
Z_G11_8	17/03/2021 (4)	011-8	370	10	90.3	0.26000	10.16467	0.07795976	0.06976	0.0016	0.15396	604.81	4.58	580.8	58.5	1.94					
Z_G11_9	17/03/2021 (4)	011-9	369	10	90.2	0.26000	10.15382	0.07731735	0.06942	0.0017	0.205	605.5	4.4	617.7	56.7	1.07					
Z_G11_10	17/03/2021 (4)	011-10	362	10	85.9	0.27000	10.17376	0.07884158	0.0594	0.0016	0.15307	607.61	4.58	580.8	58.5	1.42					
Z_G11_11	17/03/2021 (4)	011-11	356	8.9	85.6	0.26000	10.16607	0.07846128	0.06976	0.0016	0.40729	609.19	4.75	639.7	59.9	4.77					
Z_G11_12	17/03/2021 (4)	011-12	362	10	85.9	0.27000	10.17376	0.07884158	0.0594	0.0016	0.15307	607.61	4.58	580.8	58.5	1.42					
Z_G11_13	17/03/2021 (4)	011-13	357	11	85.7	0.26000	10.16607	0.07846128	0.06976	0.0016	0.40729	609.19	4.75	639.7	59.9	4.77					
Z_G11_14	17/03/2021 (4)	011-14	368	10	84.4	0.26000	10.09285	0.07639928	0.06929	0.0016	0.28601	603.5	4.58	623.7	56.9	3.34					
Z_G11_15	18/03/2021 (5)	011-15	349	11	83.7	0.26000	10.11322	0.0777	0.0598	0.0016	0.1704	607.96	4.46	623.7	60.5	2.35					
Z_G11_16	18/03/2021 (5)	011-16	352	8.4	84.6	0.26000	10.09991	0.07665194	0.06966	0.0017	0.18646	608.6	4.46	595.4	58	-2.11					
Z_G11_17	18/03/2021 (5)	011-17	352.4	9.5	83.7	0.26000	10.09991	0.07542629	0.06918	0.0017	0.36568	608.84	4.34	609.1	61.1	0.04					
Z_P91500v_1	17/03/2021 (4)	pk9-1	595	12	52.17	0.93000	18.71608	0.1541283	0.05295	0.0015	0.12496	3155.5	2.69	325.6	64.3	-3.06					
Z_P91500v_2	17/03/2021 (4)	pk9-2	602	22	58.5	2.00000	18.6428	0.1540849	0.05237	0.0014	0.10869	3156.83	2.75	305.6	61	-12.05					
Z_P91500v_3	17/03/2021 (4)	pk9-3	597	10	56.1	0.93000	18.71608	0.1541283	0.05295	0.0015	0.12496	3155.5	2.69	325.6	64.3	-3.06					
Z_P91500v_4	17/03/2021 (4)	pk9-4	597	10	56.1	0.93000	18.71608	0.1541283	0.05295	0.0015	0.12496	3155.5	2.69	325.6	64.3	-3.06					
Z_P91500v_5	17/03/2021 (4)	pk9-5	597	10	56.1	0.93000	18.71608	0.1541283	0.05295	0.0015	0.12496	3155.5	2.69	325.6	64.3	-3.06					
Z_P91500v_6	17/03/2021 (4)	pk9-6	597	10	56.1	0.93000	18.71608	0.1541283	0.05295	0.0015	0.12496	3155.5	2.69	325.6	64.3	-3.06					
Z_P91500v_7	17/03/2021 (4)	pk9-7	597	10	56.1	0.93000	18.71608	0.1541283	0.05295	0.0015	0.12496	3155.5	2.69	325.6	64.3	-3.06					
Z_P91500v_8	17/03/2021 (4)	pk9-8	597	10	56.1	0.93000	18.71608	0.1541283	0.05295	0.0015	0.12496	3155.5	2.69	325.6	64.3	-3.06					
Z_P91500v_9	17/03/2021 (4)	pk9-9	597	10	56.1	0.93000	18.71608	0.1541283	0.05295	0.0015	0.12496	3155.5	2.69	325.6	64.3	-3.06					
Z_P91500v_10	17/03/2021 (4)	pk9-10	593	35	145.4	5.30000	18.68081	0.1494666	0.05413	0.0014	0.2191	3154.04	1.57	375.5	58.2	11.33					

Curtin University Geochronology center	Sample ID	Analysis date-time	Spot ID	Lu/Hf isotopic Ratios from 2016							
				Lu/Hf (217HF)							





# Appendix D3.1 continued.

Department/University/Institution	Campus ID	Academic Area	SAP ID	2023-24	2024-25	2025-26	2026-27	2027-28	2028-29	2029-30	2030-31	2031-32	2032-33	2033-34	2034-35	2035-36	2036-37	2037-38	2038-39	2039-40	2040-41	2041-42	2042-43	2043-44	2044-45	2045-46	2046-47	2047-48	2048-49	2049-50	2050-51	2051-52	2052-53	2053-54	2054-55	2055-56	2056-57	2057-58	2058-59	2059-60	2060-61	2061-62	2062-63	2063-64	2064-65	2065-66	2066-67	2067-68	2068-69	2069-70	2070-71	2071-72	2072-73	2073-74	2074-75	2075-76	2076-77	2077-78	2078-79	2079-80	2080-81	2081-82	2082-83	2083-84	2084-85	2085-86	2086-87	2087-88	2088-89	2089-90	2090-91	2091-92	2092-93	2093-94	2094-95	2095-96	2096-97	2097-98	2098-99	2099-00	2100-01	2101-02	2102-03	2103-04	2104-05	2105-06	2106-07	2107-08	2108-09	2109-10	2110-11	2111-12	2112-13	2113-14	2114-15	2115-16	2116-17	2117-18	2118-19	2119-20	2120-21	2121-22	2122-23	2123-24	2124-25	2125-26	2126-27	2127-28	2128-29	2129-30	2130-31	2131-32	2132-33	2133-34	2134-35	2135-36	2136-37	2137-38	2138-39	2139-40	2140-41	2141-42	2142-43	2143-44	2144-45	2145-46	2146-47	2147-48	2148-49	2149-50	2150-51	2151-52	2152-53	2153-54	2154-55	2155-56	2156-57	2157-58	2158-59	2159-60	2160-61	2161-62	2162-63	2163-64	2164-65	2165-66	2166-67	2167-68	2168-69	2169-70	2170-71	2171-72	2172-73	2173-74	2174-75	2175-76	2176-77	2177-78	2178-79	2179-80	2180-81	2181-82	2182-83	2183-84	2184-85	2185-86	2186-87	2187-88	2188-89	2189-90	2190-91	2191-92	2192-93	2193-94	2194-95	2195-96	2196-97	2197-98	2198-99	2199-00	2200-01	2201-02	2202-03	2203-04	2204-05	2205-06	2206-07	2207-08	2208-09	2209-10	2210-11	2211-12	2212-13	2213-14	2214-15	2215-16	2216-17	2217-18	2218-19	2219-20	2220-21	2221-22	2222-23	2223-24	2224-25	2225-26	2226-27	2227-28	2228-29	2229-30	2230-31	2231-32	2232-33	2233-34	2234-35	2235-36	2236-37	2237-38	2238-39	2239-40	2240-41	2241-42	2242-43	2243-44	2244-45	2245-46	2246-47	2247-48	2248-49	2249-50	2250-51	2251-52	2252-53	2253-54	2254-55	2255-56	2256-57	2257-58	2258-59	2259-60	2260-61	2261-62	2262-63	2263-64	2264-65	2265-66	2266-67	2267-68	2268-69	2269-70	2270-71	2271-72	2272-73	2273-74	2274-75	2275-76	2276-77	2277-78	2278-79	2279-80	2280-81	2281-82	2282-83	2283-84	2284-85	2285-86	2286-87	2287-88	2288-89	2289-90	2290-91	2291-92	2292-93	2293-94	2294-95	2295-96	2296-97	2297-98	2298-99	2299-00	2300-01	2301-02	2302-03	2303-04	2304-05	2305-06	2306-07	2307-08	2308-09	2309-10	2310-11	2311-12	2312-13	2313-14	2314-15	2315-16	2316-17	2317-18	2318-19	2319-20	2320-21	2321-22	2322-23	2323-24	2324-25	2325-26	2326-27	2327-28	2328-29	2329-30	2330-31	2331-32	2332-33	2333-34	2334-35	2335-36	2336-37	2337-38	2338-39	2339-40	2340-41	2341-42	2342-43	2343-44	2344-45	2345-46	2346-47	2347-48	2348-49	2349-50	2350-51	2351-52	2352-53	2353-54	2354-55	2355-56	2356-57	2357-58	2358-59	2359-60	2360-61	2361-62	2362-63	2363-64	2364-65	2365-66	2366-67	2367-68	2368-69	2369-70	2370-71	2371-72	2372-73	2373-74	2374-75	2375-76	2376-77	2377-78	2378-79	2379-80	2380-81	2381-82	2382-83	2383-84	2384-85	2385-86	2386-87	2387-88	2388-89	2389-90	2390-91	2391-92	2392-93	2393-94	2394-95	2395-96	2396-97	2397-98	2398-99	2399-00	2400-01	2401-02	2402-03	2403-04	2404-05	2405-06	2406-07	2407-08	2408-09	2409-10	2410-11	2411-12	2412-13	2413-14	2414-15	2415-16	2416-17	2417-18	2418-19	2419-20	2420-21	2421-22	2422-23	2423-24	2424-25	2425-26	2426-27	2427-28	2428-29	2429-30	2430-31	2431-32	2432-33	2433-34	2434-35	2435-36	2436-37	2437-38	2438-39	2439-40	2440-41	2441-42	2442-43	2443-44	2444-45	2445-46	2446-47	2447-48	2448-49	2449-50	2450-51	2451-52	2452-53	2453-54	2454-55	2455-56	2456-57	2457-58	2458-59	2459-60	2460-61	2461-62	2462-63	2463-64	2464-65	2465-66	2466-67	2467-68	2468-69	2469-70	2470-71	2471-72	2472-73	2473-74	2474-75	2475-76	2476-77	2477-78	2478-79	2479-80	2480-81	2481-82	2482-83	2483-84	2484-85	2485-86	2486-87	2487-88	2488-89	2489-90	2490-91	2491-92	2492-93	2493-94	2494-95	2495-96	2496-97	2497-98	2498-99	2499-00	2500-01	2501-02	2502-03	2503-04	2504-05	2505-06	2506-07	2507-08	2508-09	2509-10	2510-11	2511-12	2512-13	2513-14	2514-15	2515-16	2516-17	2517-18	2518-19	2519-20	2520-21	2521-22	2522-23	2523-24	2524-25	2525-26	2526-27	2527-28	2528-29	2529-30	2530-31	2531-32	2532-33	2533-34	2534-35	2535-36	2536-37	2537-38	2538-39	2539-40	2540-41	2541-42	2542-43	2543-44	2544-45	2545-46	2546-47	2547-48	2548-49	2549-50	2550-51	2551-52	2552-53	2553-54	2554-55	2555-56	2556-57	2557-58	2558-59	2559-60	2560-61	2561-62	2562-63	2563-64	2564-65	2565-66	2566-67	2567-68	2568-69	2569-70	2570-71	2571-72	2572-73	2573-74	2574-75	2575-76	2576-77	2577-78	2578-79	2579-80	2580-81	2581-82	2582-83	2583-84	2584-85	2585-86	2586-87	2587-88	2588-89	2589-90	2590-91	2591-92	2592-93	2593-94	2594-95	2595-96	2596-97	2597-98	2598-99	2599-00	2600-01	2601-02	2602-03	2603-04	2604-05	2605-06	2606-07	2607-08	2608-09	2609-10	2610-11	2611-12	2612-13	2613-14	2614-15	2615-16	2616-17	2617-18	2618-19	2619-20	2620-21	2621-22	2622-23	2623-24	2624-25	2625-26	2626-27	2627-28	2628-29	2629-30	2630-31	2631-32	2632-33	2633-34	2634-35	2635-36	2636-37	2637-38	2638-39	2639-40	2640-41	2641-42	2642-43	2643-44	2644-45	2645-46	2646-47	2647-48	2648-49	2649-50	2650-51	2651-52	2652-53	2653-54	2654-55	2655-56	2656-57	2657-58	2658-59	2659-60	2660-61	2661-62	2662-63	2663-64	2664-65	2665-66	2666-67	2667-68	2668-69	2669-70	2670-71	2671-72	2672-73	2673-74	2674-75	2675-76	2676-77	2677-78	2678-79	2679-80	2680-81	2681-82	2682-83	2683-84	2684-85	2685-86	2686-87	2687-88	2688-89	2689-90	2690-91	2691-92	2692-93	2693-94	2694-95	2695-96	2696-97	2697-98	2698-99	2699-00	2700-01	2701-02	2702-03	2703-04	2704-05	2705-06	2706-07	2707-08	2708-09	2709-10	2710-11	2711-12	2712-13	2713-14	2714-15	2715-16	2716-17	2717-18	2718-19	2719-20	2720-21	2721-22	2722-23	2723-24	2724-25	2725-26	2726-27	2727-28	2728-29	2729-30	2730-31	2731-32	2732-33	2733-34	2734-35	2735-36	2736-37	2737-38	2738-39	2739-40	2740-41	2741-42	2742-43	2743-44	2744-45	2745-46	2746-47	2747-48	2748-49	2749-50	2750-51	2751-52	2752-53	2753-54	2754-55	2755-56	2756-57	2757-58	2758-59	2759-60	2760-61	2761-62	2762-63	2763-64	2764-65	2765-66	2766-67	2767-68	2768-69	2769-70	2770-71	2771-72	2772-73	2773-74	2774-75	2775-76	2776-77	2777-78	2778-79	2779-80	2780-81	2781-82	2782-83	2783-84	2784-85	2785-86	2786-87	2787-88
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# Appendix D3.1 continued.

Analysis	Measurement ratios										Apparent ages (Ma)				Performed ages		Discord	Analysis Notes
	U		Th		207Pb		206Pb		208Pb		207Pb		206Pb		Best age (Ma)	1σ (Ma)		
	204Pb	206Pb	232Th	230Th	207Pb	206Pb	208Pb	206Pb	207Pb	206Pb	208Pb	207Pb	206Pb					
-218M33 Spot 9	174.54	8394.63	2.14	2.567	0.067	0.2229	0.0409	0.86	1291.3	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.78% used for MCA calculation	
-218M33 Spot 97	63.30	150032.50	2.01	2.567	0.072	0.231	0.0409	0.81	1295.2	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	0.37% used for MCA calculation	
-218M33 Spot 20	202.67	9430.12	2.16	2.567	0.072	0.231	0.0409	0.81	1295.2	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.28% used for MCA calculation	
-218M33 Spot 28	98.76	105470.62	4.21	2.7948	0.11	0.2267	0.0505	0.9	1354.1	34.1	1317.2	49.0	1411.0	66.8	1413.0	49.176	0.17%	
-218M33 Spot 29	125.49	77054.48	1.56	2.567	0.067	0.2229	0.0409	0.86	1291.3	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.13%	
-218M33 Spot 39	98.76	105470.62	4.21	2.7948	0.11	0.2267	0.0505	0.9	1354.1	34.1	1317.2	49.0	1411.0	66.8	1413.0	49.176	-0.13%	
-218M33 Spot 40	134.12	266124.70	1.67	2.567	0.072	0.231	0.0409	0.81	1295.2	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.13%	
-218M33 Spot 41	134.12	266124.70	1.67	2.567	0.072	0.231	0.0409	0.81	1295.2	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.13%	
-218M33 Spot 117	464.08	366707.11	3.73	2.9741	0.0897	0.2897	0.0622	0.88	1401.2	22.3	1385.1	31.6	1451.3	47.8	1451.3	31.566	1.70%	
-218M33 Spot 185	807.62	79797.47	30.77	3.1428	0.079	0.2229	0.0409	0.7	1431.2	28.3	1451.3	31.6	1451.3	47.8	1451.3	31.566	-0.94%	
-218M33 Spot 188	69.49	14741.17	4.95	3.1562	0.062	0.231	0.0506	0.82	1465.9	19.6	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.81%	
-218M33 Spot 193	282.28	62378.02	2.34	2.568	0.066	0.2472	0.0509	0.83	1402.9	21.0	1443.9	28.2	1443.9	38.4	1443.9	38.224	-0.46%	
-218M33 Spot 200	142.40	14741.17	2.15	3.0139	0.078	0.241	0.0503	0.85	1405.9	19.3	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.38%	
-218M33 Spot 201	242.49	77154.74	2.15	3.0139	0.078	0.241	0.0503	0.85	1405.9	19.3	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.38%	
-218M33 Spot 202	307.88	90681.88	2.38	3.1177	0.079	0.2404	0.0505	0.77	1371.2	18.2	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.80%	
-218M33 Spot 240	221.89	62404.51	1.16	3.1306	0.064	0.2503	0.0505	0.72	1402.7	22.7	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.88%	
-218M33 Spot 242	125.49	41277.08	1.56	2.567	0.067	0.2229	0.0409	0.86	1291.3	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.80%	
-218M33 Spot 243	103.20	72460.99	2.14	2.2048	0.081	0.2509	0.0505	0.73	1403.8	19.6	1464.4	28.2	1443.9	38.4	1443.9	38.224	-1.00%	
-218M33 Spot 244	49.63	18743.12	2.14	3.1177	0.069	0.2503	0.0505	0.77	1403.8	19.6	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.47%	
-218M33 Spot 248	47.05	28732.49	1.13	3.2344	0.11	0.258	0.0622	0.74	1404.4	25.9	1479.6	31.2	1465.0	30.7	1440	31.364	-1.41%	
-218M33 Spot 249	47.05	28732.49	1.13	3.2344	0.11	0.258	0.0622	0.74	1404.4	25.9	1479.6	31.2	1465.0	30.7	1440	31.364	-1.41%	
-218M33 Spot 250	65.34	27993.13	1.13	3.1079	0.099	0.2475	0.0504	0.73	1444.8	23.9	1426.6	27.4	1441.6	40.9	1441.6	27.44	0.82%	
-218M33 Spot 251	65.34	27993.13	1.13	3.1079	0.099	0.2475	0.0504	0.73	1444.8	23.9	1426.6	27.4	1441.6	40.9	1441.6	27.44	0.82%	
-218M33 Spot 252	282.28	62378.02	2.34	2.568	0.066	0.2472	0.0509	0.83	1402.9	21.0	1443.9	28.2	1443.9	38.4	1443.9	38.224	-0.46%	
-218M33 Spot 253	125.49	41277.08	1.56	2.567	0.067	0.2229	0.0409	0.86	1291.3	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.80%	
-218M33 Spot 254	242.49	77154.74	2.15	3.0139	0.078	0.241	0.0503	0.85	1405.9	19.3	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.38%	
-218M33 Spot 255	307.88	90681.88	2.38	3.1177	0.079	0.2404	0.0505	0.77	1371.2	18.2	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.80%	
-218M33 Spot 256	221.89	62404.51	1.16	3.1306	0.064	0.2503	0.0505	0.72	1402.7	22.7	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.88%	
-218M33 Spot 257	125.49	41277.08	1.56	2.567	0.067	0.2229	0.0409	0.86	1291.3	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.80%	
-218M33 Spot 258	103.20	72460.99	2.14	2.2048	0.081	0.2509	0.0505	0.73	1403.8	19.6	1464.4	28.2	1443.9	38.4	1443.9	38.224	-1.00%	
-218M33 Spot 259	49.63	18743.12	2.14	3.1177	0.069	0.2503	0.0505	0.77	1403.8	19.6	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.47%	
-218M33 Spot 260	47.05	28732.49	1.13	3.2344	0.11	0.258	0.0622	0.74	1404.4	25.9	1479.6	31.2	1465.0	30.7	1440	31.364	-1.41%	
-218M33 Spot 261	47.05	28732.49	1.13	3.2344	0.11	0.258	0.0622	0.74	1404.4	25.9	1479.6	31.2	1465.0	30.7	1440	31.364	-1.41%	
-218M33 Spot 262	65.34	27993.13	1.13	3.1079	0.099	0.2475	0.0504	0.73	1444.8	23.9	1426.6	27.4	1441.6	40.9	1441.6	27.44	0.82%	
-218M33 Spot 263	65.34	27993.13	1.13	3.1079	0.099	0.2475	0.0504	0.73	1444.8	23.9	1426.6	27.4	1441.6	40.9	1441.6	27.44	0.82%	
-218M33 Spot 264	282.28	62378.02	2.34	2.568	0.066	0.2472	0.0509	0.83	1402.9	21.0	1443.9	28.2	1443.9	38.4	1443.9	38.224	-0.46%	
-218M33 Spot 265	125.49	41277.08	1.56	2.567	0.067	0.2229	0.0409	0.86	1291.3	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.80%	
-218M33 Spot 266	242.49	77154.74	2.15	3.0139	0.078	0.241	0.0503	0.85	1405.9	19.3	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.38%	
-218M33 Spot 267	307.88	90681.88	2.38	3.1177	0.079	0.2404	0.0505	0.77	1371.2	18.2	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.80%	
-218M33 Spot 268	221.89	62404.51	1.16	3.1306	0.064	0.2503	0.0505	0.72	1402.7	22.7	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.88%	
-218M33 Spot 269	125.49	41277.08	1.56	2.567	0.067	0.2229	0.0409	0.86	1291.3	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.80%	
-218M33 Spot 270	103.20	72460.99	2.14	2.2048	0.081	0.2509	0.0505	0.73	1403.8	19.6	1464.4	28.2	1443.9	38.4	1443.9	38.224	-1.00%	
-218M33 Spot 271	49.63	18743.12	2.14	3.1177	0.069	0.2503	0.0505	0.77	1403.8	19.6	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.47%	
-218M33 Spot 272	47.05	28732.49	1.13	3.2344	0.11	0.258	0.0622	0.74	1404.4	25.9	1479.6	31.2	1465.0	30.7	1440	31.364	-1.41%	
-218M33 Spot 273	47.05	28732.49	1.13	3.2344	0.11	0.258	0.0622	0.74	1404.4	25.9	1479.6	31.2	1465.0	30.7	1440	31.364	-1.41%	
-218M33 Spot 274	65.34	27993.13	1.13	3.1079	0.099	0.2475	0.0504	0.73	1444.8	23.9	1426.6	27.4	1441.6	40.9	1441.6	27.44	0.82%	
-218M33 Spot 275	65.34	27993.13	1.13	3.1079	0.099	0.2475	0.0504	0.73	1444.8	23.9	1426.6	27.4	1441.6	40.9	1441.6	27.44	0.82%	
-218M33 Spot 276	282.28	62378.02	2.34	2.568	0.066	0.2472	0.0509	0.83	1402.9	21.0	1443.9	28.2	1443.9	38.4	1443.9	38.224	-0.46%	
-218M33 Spot 277	125.49	41277.08	1.56	2.567	0.067	0.2229	0.0409	0.86	1291.3	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.80%	
-218M33 Spot 278	242.49	77154.74	2.15	3.0139	0.078	0.241	0.0503	0.85	1405.9	19.3	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.38%	
-218M33 Spot 279	307.88	90681.88	2.38	3.1177	0.079	0.2404	0.0505	0.77	1371.2	18.2	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.80%	
-218M33 Spot 280	221.89	62404.51	1.16	3.1306	0.064	0.2503	0.0505	0.72	1402.7	22.7	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.88%	
-218M33 Spot 281	125.49	41277.08	1.56	2.567	0.067	0.2229	0.0409	0.86	1291.3	18.7	1297.2	25.3	1281.5	36.6	1281.5	25.284	-0.80%	
-218M33 Spot 282	103.20	72460.99	2.14	2.2048	0.081	0.2509	0.0505	0.73	1403.8	19.6	1464.4	28.2	1443.9	38.4	1443.9	38.224	-1.00%	
-218M33 Spot 283	49.63	18743.12	2.14	3.1177	0.069	0.2503	0.0505	0.77	1403.8	19.6	1464.4	28.2	1443.9	38.4	1443.9	38.224	-0.47%	
-218M33 Spot 284	47.05	28732.49	1.13	3.2344	0.11													



# Appendix D3.2 continued.

Curtin University Geochronology center		Lu-Hf Isotopic Ratios from lollite							
Spot ID	178Hf/177Hf	±2σ INT	176Hf/177Hf	±2σ INT	(2% External ±)	176Lu/177Hf	±2σ INT	176Yb/177Hf	±2σ INT
			0.282785			0.03360 Bouvier et al., 2008 EPL-273			
			0.283250			0.03840 Griffin et al., 2002, Lithos-237			
Run 1: Standard Results - Lu/Hf on Nu Plasma II - samples 10DTB19, 12DTB19, 14DTB19 - MudTank Primary Standard - errors propagated within run and are similar to long term running averages (see Spencer et al., 2020).									
FC - 1	1.4675	0.00014	0.282202	0.000045	0.00568904	0.001822	0.000031	0.0775	0.0012
FC - 2	1.46752	0.00013	0.28222	0.000043	0.0056874	0.0015052	0.000017	0.06502	0.0004
FC - 3	1.46762	0.00013	0.282169	0.000044	0.00568738	0.0010627	0.0000046	0.04379	0.00015
FC - 4	1.46753	0.00015	0.282286	0.000036	0.00568172	0.001807	0.0000031	0.07137	0.00068
FC - 5	1.46747	0.00013	0.28227	0.000033	0.0056784	0.001857	0.0000037	0.07218	0.00087
FC - 6	1.46751	0.00015	0.282185	0.000036	0.0056797	0.0016159	0.000003	0.06984	0.00049
FC - 7	1.46741	0.00014	0.282195	0.000033	0.0056769	0.001255	0.0000015	0.05191	0.00034
FC - 8	1.46733	0.00014	0.282146	0.000036	0.00567892	0.00085855	0.0000059	0.03471	0.00026
FC - 9	1.46744	0.00016	0.28219	0.000044	0.0056878	0.00141074	0.0000059	0.05874	0.00036
FC - 10	1.46738	0.00014	0.282257	0.00003	0.00567514	0.0016104	0.0000063	0.06621	0.00067
FC - 11	1.46729	0.00017	0.282203	0.000041	0.00568506	0.001545	0.0000065	0.06521	0.00056
FC - 12	1.46735	0.00015	0.282252	0.000037	0.00568204	0.001637	0.0000019	0.06753	0.00068
G11 - 1	1.46762	0.00015	0.281992	0.000054	0.00569384	0.00034026	0.0000096	0.013194	0.00081
G11 - 2	1.46768	0.00013	0.281963	0.000051	0.00569026	0.00033693	0.0000066	0.012847	0.00078
G11 - 3	1.46768	0.00015	0.282029	0.000049	0.00568958	0.00033294	0.0000084	0.012669	0.00074
G11 - 4	1.46763	0.00015	0.281994	0.000049	0.00568888	0.00032519	0.0000007	0.012306	0.00077
G11 - 5	1.46753	0.00015	0.282006	0.000049	0.00568912	0.00032549	0.0000081	0.012247	0.00068
G11 - 6	1.46751	0.00015	0.281969	0.000052	0.00569138	0.00033002	0.0000072	0.012347	0.00082
G11 - 7	1.46751	0.00012	0.282029	0.000046	0.00568658	0.00032936	0.0000074	0.012266	0.00079
G11 - 8	1.46749	0.00014	0.282079	0.000052	0.00569358	0.00032847	0.0000064	0.012166	0.00074
G11 - 9	1.46754	0.00015	0.28205	0.000045	0.005686	0.00032275	0.0000065	0.011949	0.00068
G11 - 10	1.46742	0.00012	0.282006	0.00005	0.00569012	0.00032044	0.0000085	0.011804	0.00062
G11 - 11	1.46752	0.00015	0.282021	0.000041	0.00568142	0.00032666	0.0000074	0.011931	0.00066
G11 - 12	1.46747	0.00016	0.282	0.000043	0.005683	0.0003219	0.0000007	0.011817	0.00066
G11 - 13	1.46752	0.00017	0.281998	0.00004	0.00567996	0.00032491	0.0000073	0.011934	0.00063
G11 - 14	1.46742	0.00014	0.282034	0.000055	0.00569568	0.00032222	0.0000008	0.011824	0.00065
G11 - 15	1.46749	0.00017	0.281995	0.000043	0.0056829	0.00031295	0.0000007	0.011472	0.00069
G11 - 16	1.46743	0.00013	0.282057	0.000056	0.00569714	0.0003197	0.0000063	0.011659	0.00067
G11 - 17	1.46752	0.00015	0.282026	0.000047	0.00568752	0.00031975	0.0000064	0.011692	0.00078
G11 - 18	1.46748	0.00017	0.282015	0.000043	0.0056833	0.00032387	0.0000073	0.011796	0.00071
G11 - 19	1.4674	0.00016	0.281985	0.000045	0.0056847	0.00032148	0.0000075	0.011776	0.00069
G11 - 20	1.46747	0.00016	0.282003	0.000039	0.00567906	0.00031987	0.0000079	0.011697	0.00068
MUD - 1	1.4676	0.00013	0.282529	0.000038	0.00568858	0.00002452	0.0000039	0.001217	0.00015
MUD - 2	1.46754	0.00015	0.282497	0.000044	0.00569394	0.00002601	0.0000051	0.001252	0.00024
MUD - 3	1.46759	0.00015	0.282482	0.000046	0.00569564	0.00002451	0.0000043	0.001205	0.00016
MUD - 4	1.46765	0.00014	0.282521	0.000035	0.00568542	0.00002353	0.0000045	0.001148	0.00015
MUD - 5	1.4676	0.00013	0.282495	0.000035	0.0056849	0.0000246	0.0000035	0.001171	0.00019
MUD - 6	1.46754	0.00012	0.282504	0.000042	0.00569208	0.00002843	0.0000044	0.001364	0.00016
MUD - 7	1.46759	0.00011	0.282492	0.000041	0.00569084	0.00002497	0.0000055	0.001194	0.00023
MUD - 8	1.46751	0.00014	0.282495	0.000029	0.0056789	0.00002584	0.0000046	0.001203	0.00013
MUD - 9	1.46749	0.00013	0.282524	0.000035	0.00568548	0.00002834	0.0000047	0.001374	0.00014
MUD - 10	1.46752	0.00015	0.282522	0.000027	0.00567744	0.00002445	0.0000057	0.001158	0.00015
MUD - 11	1.4675	0.00018	0.282487	0.000042	0.00569174	0.00002426	0.0000005	0.001153	0.00014
MUD - 12	1.46749	0.00016	0.282508	0.00003	0.00568016	0.00002417	0.0000004	0.001147	0.00013
MUD - 14	1.4674	0.00015	0.282494	0.000037	0.00568688	0.00002309	0.0000044	0.001115	0.00002
MUD - 15	1.46743	0.00016	0.282528	0.000034	0.00568456	0.00002751	0.0000041	0.001283	0.00014
MUD - 16	1.46753	0.00013	0.28251	0.000037	0.0056872	0.00002586	0.0000057	0.001195	0.00023
MUD - 17	1.46747	0.00016	0.282502	0.000042	0.00569204	0.00002524	0.0000049	0.001166	0.00015
MUD - 18	1.46748	0.00015	0.282485	0.000036	0.0056857	0.00002692	0.0000005	0.001234	0.00017
MUD - 19	1.46735	0.00012	0.282514	0.000038	0.00568828	0.00002109	0.0000047	0.000979	0.00012
MUD - 20	1.46736	0.00015	0.282514	0.000037	0.00568728	0.00002282	0.0000039	0.001053	0.00001
PLES - 1	1.46762	0.00018	0.282458	0.000034	0.00568316	0.0001439	0.0000046	0.008489	0.00054
PLES - 2	1.46762	0.00014	0.282465	0.000033	0.0056823	0.00013141	0.0000038	0.007894	0.00054
PLES - 3	1.4676	0.00016	0.282458	0.000042	0.00569116	0.00013048	0.0000031	0.007741	0.00056
PLES - 4	1.46758	0.00016	0.282489	0.000039	0.00568878	0.00012936	0.0000044	0.007611	0.00053
PLES - 5	1.46748	0.00016	0.282503	0.000037	0.00568706	0.00013587	0.0000039	0.007861	0.00043
PLES - 6	1.46742	0.00018	0.282475	0.000036	0.0056855	0.00013753	0.0000045	0.007855	0.00045
PLES - 7	1.46745	0.00013	0.282482	0.000032	0.00568164	0.00011909	0.0000077	0.006543	0.00024
PLES - 8	1.46743	0.00015	0.282468	0.000033	0.00568236	0.00012871	0.0000086	0.00704	0.00018
PLES - 9	1.46748	0.00014	0.282462	0.000034	0.00568324	0.00012964	0.0000085	0.007123	0.00027
PLES - 10	1.46743	0.00014	0.282493	0.000042	0.00569186	0.00013199	0.0000047	0.007285	0.00037
PLES - 11	1.46739	0.00011	0.282464	0.000039	0.00568828	0.00011219	0.0000058	0.006106	0.00026
R33 - 1	1.46768	0.00013	0.282741	0.000042	0.00569682	0.0013344	0.0000085	0.05549	0.00079
R33 - 2	1.46768	0.00016	0.282729	0.000037	0.00569158	0.0013444	0.0000073	0.05253	0.00056
R33 - 3	1.4676	0.00015	0.282805	0.000034	0.0056901	0.001936	0.0000021	0.0764	0.0018
R33 - 4	1.46764	0.00014	0.282743	0.000033	0.00568786	0.001235	0.0000018	0.04544	0.00053
R33 - 5	1.4676	0.00017	0.282795	0.000046	0.0057019	0.003598	0.0000033	0.15253	0.00055
R33 - 6	1.4676	0.00015	0.282755	0.000038	0.0056931	0.001788	0.0000016	0.0727	0.0011
R33 - 7	1.46754	0.00015	0.282806	0.000031	0.00568712	0.00209363	0.0000082	0.08579	0.00053
R33 - 8	1.46746	0.00013	0.282772	0.000039	0.00569444	0.002142	0.0000046	0.0895	0.0025
R33 - 11	1.4675	0.00015	0.282779	0.000043	0.00569588	0.001098	0.0000089	0.0427	0.0033
R33 - 12	1.46747	0.00014	0.282737	0.000039	0.00569374	0.0010988	0.0000054	0.04337	0.00043
R33 - 13	1.46741	0.00016	0.28276	0.000038	0.0056932	0.002432	0.0000026	0.0979	0.0016
R33 - 14	1.46745	0.00014	0.282773	0.000045	0.00570046	0.001081	0.0000016	0.04231	0.00093
R33 - 15	1.46748	0.00011	0.282757	0.000042	0.00569714	0.001535	0.0000091	0.0626	0.0043
R33 - 16	1.46749	0.00011	0.282742	0.000053	0.00570784	0.002291	0.0000044	0.0943	0.0013
R33 - 17	1.46744	0.00014	0.282707	0.000038	0.00569214	0.00132	0.000004	0.0508	0.0015
R33 - 18	1.46743	0.00017	0.282732	0.000038	0.00569264	0.0011112	0.0000059	0.04263	0.00017
R33 - 19	1.46747	0.00017	0.282753	0.00004	0.00569506	0.002054	0.0000011	0.08185	0.00029
R33 - 20	1.46739	0.00013	0.282744	0.000031	0.00568588	0.00135	0.0000034	0.0511	0.0012
91500 - 1	1.46751	0.00015	0.282294	0.000055	0.00570088	0.00031298	0.0000078	0.012524	0.00078
91500 - 2	1.46756	0.00015	0.282271	0.000043	0.00568842	0.00030758	0.0000073	0.012144	0.00073
91500 - 3	1.46762	0.00014	0.282258	0.000052	0.00569716	0.0003064	0.0000011	0.011946	0.00079
91500 - 4	1.46764	0.00014	0.282304	0.000058	0.00570408	0.00031624	0.0000082	0.012322	0.00059
91500 - 5	1.46774	0.00017	0.282276	0.000046	0.00569152	0.00030084	0.0000086	0.011515	0.00084
91500 - 6	1.46759	0.00016	0.282335	0.000044	0.0056907	0.00030118	0.0000091	0.011583	0.00066
91500 - 7	1.4676	0.00014	0.282326	0.000054	0.00570052	0.00030039	0.0000008	0.011477	0.00073
91500 - 8	1.46753	0.00017	0.282331	0.000056	0.00570262	0.00029342	0.0000009	0.011097	0.00086
91500 - 10	1.46752	0.00018	0.282304	0.000053	0.00569908	0.00030244	0.0000084	0.011497	0.00072
91500 - 12	1.46753	0.00013	0.282316	0.000058	0.005				



# Appendix D3.2 continued.

Curtin University Geohistory center		Lu-Hf Isotopic Ratios from lollite							
Spot ID	178Hf/177Hf	±2σ INT	176Hf/177Hf	±2σ INT	(2% External ±)	176Lu/177Hf	±2σ INT	176Yb/177Hf	±2σ INT
			0.282785 0.283250			0.03360 Bouvier et al., 2008 EPLS-273 0.03840 Griffin et al., 2002, Lithos-237			
Run 2: Standard Results - Lu/Hf on Nu Plasma II - samples 38DTB19, 84DTB19 - MudTank Primary Standard - errors propagated within run and are similar to long term running averages (see Spencer et al., 2020).									
FC1_1	1.4675	0.00013	0.282156	0.000038	0.001276	0.001276	0.000015	0.05064	0.00032
FC1_2	1.46731	0.00012	0.282196	0.000039	0.000847	0.000847	0.000014	0.03236	0.00034
FC1_3	1.46722	0.00016	0.282156	0.000043	0.000903	0.000903	0.000023	0.03438	0.00074
FC1_4	1.46726	0.00012	0.282194	0.000045	0.001273	0.001273	0.000016	0.04913	0.00043
FC1_5	1.46722	0.00015	0.282186	0.000048	0.001272	0.001272	0.000012	0.04971	0.0002
FC1_6	1.46727	0.00015	0.282137	0.000045	0.001191	0.001191	0.000018	0.04701	0.00037
FC1_7	1.46753	0.00014	0.282152	0.00004	0.001154	0.001154	0.000011	0.04459	0.00077
FC1_8	1.46751	0.000099	0.282195	0.00004	0.0011722	0.0011722	0.0000048	0.04499	0.00041
FC1_9	1.46723	0.00012	0.282145	0.000038	0.00074055	0.00074055	0.0000051	0.0276	0.00019
FC1_10	1.46716	0.00013	0.282175	0.000038	0.001176	0.001176	0.000015	0.04558	0.00031
FC1_11	1.46748	0.00014	0.282202	0.000043	0.00568704	0.001242	0.00005	0.0457	0.0016
FC1_12	1.46752	0.00015	0.282199	0.000046	0.00568998	0.001058	0.00001	0.03982	0.00014
FC1_13	1.46759	0.00012	0.282174	0.000044	0.00568748	0.000758	0.0000018	0.02708	0.00029
FC1_14	1.46749	0.00013	0.282156	0.00004	0.00568312	0.0008106	0.0000018	0.02957	0.00033
Z_G11_1	1.4675	0.00014	0.281998	0.000054	0.00028407	0.00028407	0.00000079	0.009863	0.00052
Z_G11_2	1.46735	0.00016	0.282066	0.00005	0.00028781	0.00028781	0.00000076	0.009947	0.00055
Z_G11_3	1.46726	0.00014	0.282013	0.000054	0.00028533	0.00028533	0.00000085	0.009889	0.00063
Z_G11_4	1.46733	0.00015	0.282042	0.000055	0.00028712	0.00028712	0.00000075	0.010002	0.00059
Z_G11_5	1.46739	0.00015	0.282041	0.000048	0.00028556	0.00028556	0.00000068	0.009961	0.00057
Z_G11_6	1.46734	0.00016	0.282026	0.000055	0.00028051	0.00028051	0.00000078	0.009715	0.00062
Z_G11_7	1.46752	0.00017	0.281949	0.000043	0.0002832	0.0002832	0.00000056	0.009875	0.00064
Z_G11_8	1.46754	0.00015	0.281972	0.00005	0.00028436	0.00028436	0.00000058	0.009819	0.00059
Z_G11_9	1.46762	0.00014	0.282026	0.000043	0.00028626	0.00028626	0.0000006	0.009904	0.00062
Z_G11_10	1.46745	0.00014	0.282064	0.000064	0.0002902	0.0002902	0.00000068	0.010024	0.00051
Z_G11_11	1.46745	0.00013	0.282093	0.000049	0.00028669	0.00028669	0.00000082	0.009916	0.00055
Z_G11_12	1.46716	0.00014	0.281996	0.000049	0.00028022	0.00028022	0.00000079	0.009679	0.00005
Z_G11_13	1.46715	0.00012	0.282008	0.000048	0.00028738	0.00028738	0.00000077	0.00996	0.00054
Z_G11_14	1.46729	0.00015	0.281979	0.000054	0.00028492	0.00028492	0.0000006	0.009885	0.00057
Z_G11_15	1.4672	0.00017	0.282006	0.000044	0.00028281	0.00028281	0.00000052	0.009801	0.00063
Z_MT_1	1.46751	0.00016	0.282511	0.000043	0.00002161	0.00002161	0.00000004	0.000949	0.00017
Z_MT_2	1.46736	0.00014	0.2825	0.000035	0.00002158	0.00002158	0.00000059	0.000961	0.00013
Z_MT_3	1.46733	0.00012	0.282514	0.000035	0.00001843	0.00001843	0.00000041	0.000913	0.00014
Z_MT_4	1.46733	0.00013	0.282504	0.000039	0.00001925	0.00001925	0.00000041	0.000968	0.00015
Z_MT_5	1.46726	0.00017	0.282506	0.000038	0.00002006	0.00002006	0.00000044	0.000968	0.00018
Z_MT_6	1.46732	0.00015	0.282507	0.000047	0.00002054	0.00002054	0.00000047	0.000906	0.00016
Z_MT_7	1.46725	0.00013	0.282508	0.000041	0.00002444	0.00002444	0.00000058	0.001076	0.00018
Z_MT_8	1.46726	0.00014	0.282506	0.000043	0.00002033	0.00002033	0.00000051	0.00092	0.00013
Z_MT_9	1.46735	0.00012	0.282509	0.000042	0.00002409	0.00002409	0.00000033	0.001041	0.0002
Z_MT_10	1.46756	0.00014	0.282503	0.000033	0.00002157	0.00002157	0.00000005	0.000932	0.0002
Z_MT_11	1.46759	0.0001	0.282515	0.000038	0.00002395	0.00002395	0.00000048	0.001047	0.00017
Z_MT_12	1.46737	0.00012	0.282502	0.000039	0.00002245	0.00002245	0.00000041	0.000944	0.00015
Z_MT_13	1.4674	0.00015	0.282505	0.000035	0.00002218	0.00002218	0.00000042	0.000945	0.0002
Z_MT_14	1.46725	0.00016	0.282511	0.00003	0.00002532	0.00002532	0.00000039	0.001081	0.00016
Z_MT_15	1.46723	0.00013	0.282505	0.000041	0.00002663	0.00002663	0.00000042	0.001161	0.00017
Z_MT_16	1.46733	0.00016	0.282507	0.000048	0.00002774	0.00002774	0.00000042	0.001212	0.00013
Z_MT_17	1.46723	0.00016	0.282507	0.000038	0.00003086	0.00003086	0.00000042	0.001349	0.00014
PLES_1	1.46733	0.0001	0.282503	0.000031	0.00006283	0.00006283	0.00000056	0.003264	0.00014
PLES_2	1.46732	0.00014	0.282474	0.000039	0.00019877	0.00019877	0.00000099	0.01037	0.00013
PLES_3	1.46756	0.00013	0.282474	0.000044	0.0002083	0.0002083	0.0000011	0.01094	0.00012
PLES_4	1.46727	0.0001	0.282485	0.000036	0.00021616	0.00021616	0.00000052	0.011238	0.00066
PLES_5	1.46726	0.00017	0.282502	0.000038	0.0002133	0.0002133	0.0000012	0.011044	0.00002
PLES_6	1.46738	0.00014	0.282484	0.000043	0.0001979	0.0001979	0.0000019	0.01028	0.00041
PLES_7	1.46745	0.00012	0.28247	0.000029	0.00017293	0.00017293	0.00000037	0.008842	0.00058
PLES_8	1.46757	0.00011	0.282457	0.000033	0.00016275	0.00016275	0.00000056	0.008246	0.00045
PLES_9	1.46746	0.00014	0.282471	0.000039	0.00020945	0.00020945	0.00000098	0.010899	0.00069
PLES_10	1.46725	0.00015	0.282499	0.000044	0.000214	0.000214	0.0000014	0.011142	0.00062
PLES_11	1.46721	0.00011	0.282483	0.000032	0.0002137	0.0002137	0.0000016	0.011141	0.00074
Z_R33_1	1.46742	0.00015	0.282744	0.000043	0.0011121	0.0011121	0.000003	0.03892	0.00017
Z_R33_2	1.4674	0.00013	0.282804	0.000046	0.0017781	0.0017781	0.0000079	0.06848	0.00078
Z_R33_3	1.46726	0.00015	0.282777	0.000049	0.0015899	0.0015899	0.0000099	0.05576	0.00021
Z_R33_4	1.4674	0.00017	0.282805	0.000045	0.0010866	0.0010866	0.0000091	0.03777	0.00013
Z_R33_5	1.46733	0.00017	0.282779	0.000042	0.0017139	0.0017139	0.0000078	0.06405	0.00091
Z_R33_6	1.46726	0.0001	0.282768	0.000039	0.0017211	0.0017211	0.0000093	0.05775	0.00065
Z_R33_7	1.46741	0.00012	0.282782	0.000034	0.001513	0.001513	0.0000016	0.05853	0.00088
Z_R33_8	1.46721	0.00012	0.282768	0.000045	0.0013796	0.0013796	0.0000081	0.05169	0.0006
Z_R33_9	1.46729	0.00014	0.282724	0.000042	0.001171	0.001171	0.0000011	0.042192	0.00063
Z_R33_10	1.46747	0.00013	0.282799	0.000041	0.00275	0.00275	0.0000028	0.1083	0.0011
Z_R33_11	1.46746	0.00015	0.282715	0.000044	0.0023244	0.0023244	0.0000021	0.08921	0.00066
Z_R33_12	1.46735	0.00011	0.28277	0.000033	0.001321	0.001321	0.0000018	0.04911	0.00035
Z_R33_13	1.46742	0.00015	0.282802	0.00003	0.0013815	0.0013815	0.000003	0.05045	0.00044
Z_R33_14	1.46723	0.00015	0.282787	0.000046	0.001739	0.001739	0.0000014	0.06491	0.00098
Z_R33_15	1.46712	0.00014	0.282781	0.000043	0.00206	0.00206	0.0000015	0.0783	0.0054
Z_R33_16	1.46722	0.00013	0.282736	0.000041	0.001342	0.001342	0.0000028	0.051	0.0014
Z_R33_17	1.46727	0.00015	0.282789	0.000045	0.0009841	0.0009841	0.0000027	0.03562	0.00019
Z_91500_1	1.46755	0.00015	0.282272	0.000049	0.00027618	0.00027618	0.00000075	0.009809	0.00072
Z_91500_2	1.46732	0.00018	0.282269	0.000063	0.00027121	0.00027121	0.00000085	0.009692	0.00064
Z_91500_3	1.46741	0.00016	0.282268	0.000049	0.00026224	0.00026224	0.00000081	0.009525	0.00075
Z_91500_4	1.46727	0.00013	0.282357	0.000057	0.000281	0.000281	0.00000076	0.009951	0.00063
Z_91500_5	1.46727	0.0002	0.282267	0.000067	0.00027693	0.00027693	0.00000093	0.009828	0.00051
Z_91500_6	1.4672	0.00019	0.282278	0.000046	0.0002841	0.0002841	0.000001	0.010127	0.00055
Z_91500_7	1.4673	0.00017	0.282289	0.000058	0.0002654	0.0002654	0.00000081	0.009372	0.0006
Z_91500_8	1.46733	0.0002	0.282334	0.000061	0.00027633	0.00027633	0.00000065	0.009835	0.00073
Z_91500_9	1.46741	0.00014	0.282272	0.000056	0.0002625	0.0002625	0.0000011		

# Appendix D3.2 continued.

Curtin University Geohistory center		Lu-Hf Isotopic Ratios from Iolite								
Spot ID	178Hf/177Hf	±2σ INT	176Hf/177Hf 0.282785 0.283250	±2σ INT	(% External ±)	176Lu/177Hf	±2σ INT	176yb/177Hf 0.03360 Bouvier et al., 2008 EPSL-273 0.03840 Griffin et al., 2002, Lithos-237	±2σ INT	
Run 3: Standard Results - Lu/Hf on Nu Plasma II - samples 01DTB19, 34DTB19 - MudTank Primary Standard - errors propagated within run and are similar to long term running averages (see Spencer et al., 2020).										
Z_FC1_1	1.467254	0.000068	0.282218	0.000031	0.00567536	0.0011009	0.0000076	0.04489	0.00022	
Z_FC1_2	1.46732	0.000082	0.282185	0.00002	0.0056637	0.0007049	0.0000026	0.02746	0.00017	
Z_FC1_3	1.467301	0.000073	0.282203	0.000021	0.00566506	0.0012061	0.0000024	0.04894	0.0001	
Z_FC1_4	1.467315	0.000081	0.282172	0.000026	0.00566944	0.000695	0.0000023	0.02683	0.00087	
Z_FC1_5	1.467227	0.000099	0.282356	0.000022	0.00566912	0.0030858	0.0000043	0.12093	0.00021	
Z_FC1_6	1.46718	0.000094	0.282323	0.000025	0.00567146	0.0033308	0.0000057	0.13137	0.00019	
Z_FC1_7	1.467162	0.000084	0.282262	0.000018	0.00566324	0.00222	0.000016	0.0875	0.0063	
Z_FC1_8	1.4672	0.000089	0.282265	0.000031	0.00567663	0.002	0.000012	0.0779	0.0045	
Z_FC1_9	1.467191	0.000078	0.282256	0.000028	0.00567312	0.00194	0.00001	0.0755	0.004	
Z_FC1_10	1.467243	0.00007	0.282406	0.000027	0.00567512	0.004539	0.000021	0.17949	0.00069	
Z_G11_1	1.467397	0.000092	0.28201	0.000031	0.0056712	0.00035847	0.0000007	0.013153	0.00028	
Z_G11_2	1.467297	0.000078	0.282024	0.000024	0.00566448	0.00033509	0.0000036	0.012197	0.00038	
Z_G11_3	1.46741	0.00011	0.28203	0.000024	0.0056646	0.00032499	0.0000005	0.011779	0.00034	
Z_G11_4	1.46737	0.000092	0.282	0.000027	0.005667	0.0003252	0.0000057	0.011649	0.00027	
Z_G11_5	1.46727	0.00011	0.282015	0.000029	0.0056693	0.00031846	0.0000043	0.01143	0.00023	
Z_G11_6	1.467389	0.000094	0.282025	0.000022	0.0056625	0.00032197	0.0000053	0.011541	0.00049	
Z_G11_7	1.467301	0.000081	0.281991	0.000034	0.00567382	0.00031978	0.0000055	0.011363	0.00028	
Z_G11_8	1.467242	0.000077	0.282005	0.000036	0.0056761	0.00031952	0.0000047	0.011319	0.00023	
Z_G11_9	1.46725	0.0001	0.282017	0.000027	0.00566734	0.00031515	0.0000044	0.011095	0.00037	
Z_G11_10	1.467345	0.000083	0.282015	0.000021	0.0056613	0.00031233	0.0000055	0.010985	0.00029	
Z_G11_11	1.46727	0.00012	0.282021	0.000024	0.00566442	0.00031723	0.0000005	0.011202	0.00042	
Z_G11_12	1.46724	0.00011	0.282014	0.000027	0.00566728	0.00031306	0.0000045	0.011021	0.00029	
Z_G11_13	1.467245	0.000096	0.28202	0.000028	0.0056684	0.00031188	0.0000054	0.010993	0.00021	
Z_G11_14	1.467295	0.000086	0.282012	0.000025	0.00566524	0.00032024	0.0000057	0.011242	0.00027	
Z_G11_15	1.467276	0.00009	0.282021	0.00003	0.00567042	0.00032266	0.0000043	0.011297	0.00024	
Z_G11_16	1.467168	0.000092	0.282023	0.000026	0.00566646	0.00032254	0.0000055	0.011315	0.00023	
Z_G11_17	1.467268	0.000081	0.281987	0.000029	0.00566874	0.00031169	0.0000062	0.011018	0.00017	
Z_G11_18	1.467242	0.000076	0.282008	0.00002	0.00566016	0.0003323	0.000001	0.011743	0.00032	
Z_G11_19	1.4673	0.000089	0.282008	0.000025	0.00566516	0.00034473	0.0000057	0.012138	0.00028	
Z_MT_1	1.46734	0.0001	0.282512	0.00002	0.00567024	0.00002986	0.0000023	0.0013993	0.000071	
Z_MT_2	1.467462	0.000079	0.2825	0.000031	0.005681	0.00002819	0.0000026	0.0013011	0.000051	
Z_MT_3	1.46731	0.00011	0.282489	0.000023	0.00567278	0.0000264	0.0000017	0.001255	0.000069	
Z_MT_4	1.467287	0.000076	0.282522	0.000023	0.00567344	0.00002742	0.0000015	0.001243	0.000064	
Z_MT_5	1.467339	0.000068	0.282511	0.000023	0.00567322	0.00002822	0.0000019	0.0012875	0.000064	
Z_MT_6	1.467273	0.000086	0.282519	0.000027	0.00567738	0.0000224	0.000001	0.000991	0.000045	
Z_MT_7	1.467316	0.000071	0.28249	0.000021	0.0056708	0.00001198	0.0000048	0.000475	0.00017	
Z_MT_8	1.467319	0.000088	0.282507	0.000022	0.00567214	0.000025	0.0000018	0.001341	0.000049	
Z_MT_9	1.467303	0.000079	0.282504	0.000023	0.00567308	0.00002507	0.0000022	0.001255	0.000054	
Z_MT_10	1.467322	0.000076	0.282519	0.00002	0.00567038	0.00002534	0.0000023	0.001267	0.000049	
Z_MT_11	1.467284	0.000092	0.282514	0.000017	0.00566728	0.0000147	0.0000012	0.00062	0.00053	
Z_MT_12	1.467265	0.000075	0.282502	0.000013	0.00566304	0.00001442	0.0000029	0.000566	0.00012	
Z_MT_13	1.46723	0.0001	0.282517	0.000023	0.0056734	0.00002297	0.0000019	0.0010103	0.000058	
Z_MT_14	1.467213	0.000086	0.282504	0.000024	0.00567408	0.00002398	0.0000022	0.0010762	0.000072	
Z_MT_15	1.467306	0.000081	0.282481	0.000023	0.00567262	0.00001626	0.0000035	0.000649	0.00015	
Z_MT_16	1.467232	0.000075	0.282535	0.000027	0.0056777	0.00002797	0.0000019	0.001286	0.000066	
Z_MT_17	1.467221	0.000074	0.2825	0.000023	0.005673	0.00002552	0.0000018	0.0011409	0.000059	
Z_MT_18	1.467268	0.000074	0.282511	0.000025	0.00567522	0.00001299	0.0000048	0.000508	0.00018	
Z_MT_19	1.467266	0.000095	0.282518	0.000025	0.00567536	0.00002109	0.0000074	0.000887	0.00044	
Z_Plesovice_1	1.467387	0.000097	0.282495	0.000024	0.0056739	0.0001091	0.0000012	0.000595	0.00005	
Z_Plesovice_2	1.467278	0.000088	0.282467	0.000026	0.00567534	0.0000812	0.0000032	0.000437	0.00017	
Z_Plesovice_3	1.467249	0.000067	0.282464	0.000021	0.00567028	0.0000838	0.0000031	0.000447	0.00016	
Z_Plesovice_4	1.467273	0.000063	0.282464	0.000024	0.00567328	0.0001002	0.0000024	0.000533	0.00013	
Z_Plesovice_5	1.46731	0.00011	0.282471	0.000018	0.00566742	0.0000571	0.0000072	0.0003024	0.000045	
Z_Plesovice_6	1.467245	0.000094	0.282489	0.000025	0.00567478	0.0000731	0.0000042	0.000385	0.00021	
Z_Plesovice_7	1.467275	0.000084	0.282502	0.000023	0.00567304	0.0001849	0.0000039	0.000974	0.00019	
Z_Plesovice_8	1.467275	0.000068	0.282472	0.000029	0.00567844	0.0001867	0.0000042	0.000981	0.00023	
Z_Plesovice_9	1.46725	0.000085	0.282472	0.000024	0.00567344	0.0001792	0.0000039	0.000945	0.00023	
Z_R33_1	1.467374	0.000085	0.282826	0.000025	0.00568152	0.0024426	0.0000094	0.09872	0.00015	
Z_R33_2	1.46732	0.00011	0.282838	0.000024	0.00568076	0.002598	0.000071	0.1002	0.00024	
Z_R33_3	1.46734	0.00011	0.282777	0.000028	0.00568354	0.0012962	0.0000012	0.04448	0.00014	
Z_R33_4	1.46733	0.00011	0.282795	0.00002	0.0056759	0.0019994	0.0000058	0.07938	0.00039	
Z_R33_5	1.467386	0.000072	0.282804	0.000026	0.00568208	0.002551	0.000029	0.10173	0.0009	
Z_R33_6	1.467265	0.00008	0.282776	0.000024	0.00567952	0.002188	0.000046	0.0865	0.0021	
Z_R33_7	1.467311	0.000088	0.282756	0.000031	0.00568612	0.000659	0.000012	0.02396	0.00053	
Z_R33_8	1.46732	0.0001	0.282753	0.000025	0.00568006	0.001596	0.000041	0.0613	0.0014	
Z_R33_9	1.467287	0.000087	0.282729	0.000021	0.00567558	0.00088	0.000011	0.0306	0.00043	
Z_R33_10	1.467431	0.000099	0.282703	0.000027	0.00568106	0.001	0.000017	0.03758	0.00059	
Z_R33_11	1.4673	0.000094	0.282744	0.000025	0.00567988	0.001056	0.000034	0.0376	0.0012	
Z_R33_12	1.467251	0.000094	0.282739	0.000034	0.00568878	0.0010007	0.0000082	0.03604	0.00042	
Z_R33_13	1.467281	0.000098	0.282795	0.00002	0.0056759	0.002017	0.000076	0.07	0.0032	
Z_R33_14	1.467324	0.000066	0.282777	0.000023	0.00567854	0.001965	0.00002	0.0708	0.00091	
Z_R33_15	1.46719	0.0001	0.282789	0.000027	0.00568278	0.001623	0.000038	0.0584	0.0014	
Z_R33_16	1.467291	0.000086	0.282803	0.000027	0.00568306	0.001764	0.000017	0.06417	0.00098	
Z_R33_17	1.467269	0.000087	0.282846	0.000026	0.00568292	0.0034368	0.0000074	0.13439	0.00072	
Z_R33_18	1.467273	0.000099	0.282813	0.000026	0.00568226	0.002386	0.000034	0.09017	0.00099	
Z_R33_19	1.46725	0.000086	0.282743	0.000025	0.00567986	0.001514	0.000013	0.0538	0.00063	
Z_R33_20	1.46729	0.0001	0.282763	0.000027	0.00568226	0.00237	0.000017	0.0919	0.00066	
Z_91500_1	1.467361	0.000092	0.282317	0.000031	0.00567734	0.0003918	0.0000064	0.014902	0.00046	
Z_91500_2	1.467384	0.000099	0.282302	0.000033	0.00567904	0.0003771	0.0000068	0.014198	0.00037	
Z_91500_3	1.46729	0.00011	0.282321	0.000037	0.00568342	0.00037321	0.0000091	0.013837	0.00032	
Z_91500_4	1.46728	0.00012	0.282281	0.000034	0.00567962	0.00036631	0.0000062	0.013556	0.00027	
Z_91500_5	1.467355	0.000087	0.282297	0.000038	0.00568394	0.00036535	0.0000048	0.013472	0.00023	
Z_91500_6	1.467337	0.000091	0.282301	0.00003	0.00567602	0.00036181	0.0000042	0.013352	0.00044	
Z_91500_7	1.46735	0.0001	0.282324	0.000027	0.00567348	0.00036676	0.0000047	0.013362	0.00028	
Z_91500_8	1.467356	0.000097	0.282272	0.000032	0.00567744	0.00036352	0.0000056	0.013194	0.00026	
Z_91500_9	1.467259	0.000076	0.282297	0.000031	0.00567694	0.0003571	0.0000018	0.012878	0.00003	
Z_91500_10	1.46733	0.0001	0.2823	0.000021	0.005667	0.00035975	0.0000067	0.012975	0.00023	
Z_91500_11	1.467275	0.000074	0.282287	0.00003	0.00567574	0.00036395	0.0000			

## Appendix D3.2 continued.

Curtin University Geohistory center		Lu-Hf Isotopic Ratios from Iolite										
Spot ID	178Hf/177Hf	±2σ INT	176Hf/177Hf		±2σ INT	(2% External ±)	176Lu/177Hfm		±2σ INT	176Yb/177Hf		±2σ INT
			0.282785	0.283250			0.03360	Bouvier et al., 2008 EPSL-273		0.03840	Griffin et al., 2002, Lithos-237	

Run 4: Standard Results - Lu/Hf on Nu Plasma II - sample 38DTB19 - MudTank Primary Standard, errors propagated based on long term running averages (see Spencer et al., 2020).

Z_FC1_1	1.46748	0.00014	0.282202	0.000043	0.00568704	0.001242	0.00005	0.0457	0.0016
Z_FC1_2	1.46752	0.00015	0.282199	0.000046	0.00568998	0.001058	0.00001	0.03982	0.00014
Z_FC1_3	1.46759	0.00012	0.282174	0.000044	0.00568748	0.000758	0.0000018	0.02708	0.00029
Z_FC1_4	1.46749	0.00013	0.282156	0.00004	0.00568312	0.0008106	0.0000018	0.02957	0.00033
Z_GJ1_1	1.4675	0.00012	0.28198	0.00005	0.0056896	0.00028214	0.00000076	0.009402	0.000061
Z_GJ1_2	1.46755	0.00014	0.282079	0.000048	0.00568958	0.00028185	0.00000095	0.009389	0.000072
Z_GJ1_3	1.46754	0.00012	0.28195	0.000055	0.005694	0.00028301	0.00000073	0.009457	0.000064
Z_GJ1_4	1.46752	0.00014	0.282029	0.00005	0.00569058	0.0002831	0.00000061	0.00942	0.000076
Z_GJ1_5	1.46753	0.00014	0.282049	0.000042	0.00568298	0.0002816	0.00000067	0.00929	0.000072
Z_MT_1	1.4675	0.00013	0.282512	0.000035	0.00568524	0.00001927	0.00000053	0.000823	0.000022
Z_MT_2	1.46753	0.00012	0.282496	0.000047	0.00569692	0.00002075	0.00000038	0.000912	0.000021
Z_MT_3	1.46743	0.00012	0.282493	0.000049	0.00569886	0.00002006	0.00000054	0.000858	0.000016
Z_MT_4	1.46752	0.00013	0.282531	0.000048	0.00569862	0.00002338	0.0000006	0.000993	0.000021
Z_MT_5	1.46754	0.00013	0.2825	0.000045	0.005695	0.0000216	0.00000073	0.000919	0.00002
Z_Plesovice_1	1.46751	0.00014	0.28253	0.000037	0.0056876	0.000162	0.000001	0.007983	0.000095
Z_Plesovice_2	1.46751	0.00013	0.282534	0.000037	0.00568768	0.000156	0.0000012	0.007699	0.00003
Z_Plesovice_3	1.4675	0.00012	0.28244	0.00005	0.0056988	0.00014762	0.00000086	0.007343	0.000034
Z_Plesovice_4	1.46745	0.00011	0.282419	0.000038	0.00568638	0.0001333	0.0000011	0.006817	0.000024
Z_R33_1	1.467431	0.000098	0.282768	0.000042	0.00569736	0.001047	0.000019	0.03616	0.00048
Z_R33_2	1.46755	0.00014	0.282796	0.000055	0.00571092	0.00223	0.00013	0.0848	0.00057
Z_R33_3	1.46754	0.00012	0.282771	0.000055	0.00571042	0.001826	0.000026	0.0642	0.00015
Z_R33_4	1.46758	0.00014	0.282741	0.000043	0.00569782	0.0015573	0.0000035	0.05245	0.00018
Z_R33_5	1.46752	0.00013	0.282732	0.000043	0.00569764	0.0011273	0.0000025	0.03945	0.00031
Z_91500_1	1.46752	0.00014	0.282292	0.000058	0.00570384	0.00025983	0.00000082	0.008756	0.00007
Z_91500_2	1.46767	0.00014	0.282339	0.000059	0.00570578	0.00028391	0.00000097	0.009675	0.000075
Z_91500_3	1.46758	0.00015	0.282321	0.000065	0.00571142	0.0002791	0.000001	0.009504	0.000052
Z_91500_4	1.46738	0.00015	0.282304	0.000058	0.00570408	0.0002601	0.0000011	0.008873	0.000088
Z_91500_5	1.46752	0.00014	0.282323	0.000056	0.00570246	0.0002719	0.0000012	0.009287	0.000072

## Appendix D3.2 continued.

Table 21JBM18. U-Pb geochronologic analyse University of Arizona Laserchron center										Measured ratios				Apparent ages (Ma)			
Analysis	U (ppm)	206Pb 204Pb	U/Th	207Pb ±		206Pb ±		error corr.	207Pb ±		206Pb ±		206Pb ±				
				235U	2 sig	238U	2 sig		235U	(Ma) 2 sig	238U	(Ma) 2 sig	207Pb	(Ma) 2 sig			
Standard Results - U/Pb on 7700 - Samples 10DTB19, 12DTB19, 14DTB19 - OGC primary standard for unknowns > 1500 Ma, 91500 primary standard for unknowns > 1500 Ma, errors propagated within run and are the same as long term running averages (Spencer et al., 2016)																	
FC-1	271	69015	2.21	1.98649	0.0572	0.18848	0.00422	0.78	1110.93	17.6904	1113.1	20.748	1106.6	32.76			
FC-1	1139	1088577	1.50	1.93898	0.0671	0.18735	0.00596	0.92	1094.6	21.112	1107	29.484	1070.2	24.752			
FC-1	501	58287	1.59	1.92926	0.0722	0.18461	0.00591	0.86	1091.3	22.75	1092.1	29.302	1089.6	34.762			
FC-1	463	78941	1.76	1.91642	0.0567	0.18351	0.00484	0.89	1086.82	17.9634	1086.1	24.024	1088.2	24.57			
FC-1	505	99066	1.44	1.93512	0.0716	0.18549	0.00612	0.90	1093.3	22.568	1096.9	30.212	1086.2	29.484			
FC-1	543	154574	1.73	1.96775	0.0649	0.18805	0.00549	0.89	1104.5	20.202	1110.8	27.118	1092.2	27.482			
FC-1	291	288927	1.95	1.97963	0.0641	0.18777	0.00511	0.84	1108.6	19.838	1109.3	25.298	1107.2	32.032			
FC-1	681	795728	2.25	1.97407	0.0612	0.18674	0.00504	0.87	1106.7	18.928	1103.7	24.934	1112.6	27.846			
FC-1	293	77021	1.74	1.94369	0.0439	0.18578	0.0031	0.74	1096.27	13.7774	1098.47	15.3426	1091.9	27.664			
FC-1	244	72996	1.62	1.95376	0.0508	0.18621	0.00387	0.80	1099.74	15.8886	1100.8	19.11	1097.6	28.392			
FC-1	416	289313	1.69	1.96965	0.0741	0.18807	0.00643	0.91	1105.2	23.114	1110.9	31.668	1093.9	28.392			
FC-1	512	64822	1.71	1.93836	0.0663	0.18663	0.00582	0.91	1094.4	20.93	1103.1	28.756	1077.2	25.844			
FC-1	469	125055	1.74	1.95096	0.0831	0.18704	0.00729	0.92	1098.8	26.026	1105.3	36.036	1085.8	30.394			
FC-1	230	65053	2.38	1.97984	0.0598	0.18677	0.00437	0.77	1108.7	18.564	1103.9	21.658	1118.1	34.944			
FC-1	387	96138	2.64	1.98232	0.0777	0.18834	0.00693	0.94	1109.5	24.024	1112.4	34.216	1103.9	24.388			
FC-1	358	105418	1.74	1.95244	0.0539	0.18622	0.00406	0.79	1099.29	16.8714	1100.9	20.02	1096.2	30.758			
FC-1	229	122516	1.63	1.9057	0.048	0.18305	0.0035	0.76	1083.08	15.2698	1083.62	17.3628	1082	29.848			
FC-1	305	1113304	1.90	1.95529	0.0567	0.18664	0.00463	0.85	1100.27	17.7268	1103.1	22.932	1094.6	27.846			
FC-1	505	1662322	1.41	1.94098	0.0703	0.18492	0.0061	0.91	1095.3	22.022	1093.8	30.212	1098.4	27.3			
FC-1	239	91423	1.55	1.96907	0.048	0.18805	0.0035	0.76	1104.99	14.9422	1110.8	17.29	1093.6	28.938			
FC-1	300	178604	1.82	1.94924	0.0569	0.18533	0.00393	0.72	1098.19	17.8178	1096	19.474	1102.5	36.946			
FC-1	669	278180	1.68	1.93533	0.0716	0.18444	0.00605	0.89	1093.4	22.568	1091.2	30.03	1097.8	30.758			
FC-1	168	52430	2.46	1.96506	0.0491	0.18721	0.0036	0.77	1103.62	15.3062	1106.24	17.7814	1098.4	29.12			
FC-1	291	61211	1.51	1.98762	0.0529	0.18773	0.00402	0.80	1111.31	16.3618	1109.1	19.838	1115.7	28.938			
FC-1	145	110592	2.23	1.92675	0.0478	0.18295	0.0033	0.73	1090.41	15.0878	1083.07	16.3618	1105.1	30.758			
FC-1	672	143084	1.77	1.94379	0.0568	0.18577	0.00498	0.92	1096.31	17.836	1098.4	24.57	1092.1	20.93			
FC-1	427	119544	1.85	1.94702	0.0794	0.1856	0.00709	0.93	1097.4	24.934	1097.5	35.126	1097.3	27.3			
FC-1	247	42914	1.63	1.94462	0.0572	0.18555	0.00423	0.78	1096.59	17.9452	1097.2	20.93	1095.3	33.488			
FC-1	236	79783	2.20	1.95554	0.0583	0.18543	0.00438	0.79	1100.4	18.2	1096.6	21.658	1107.8	33.306			
FC-1	1171	267399	1.38	1.94149	0.0652	0.18546	0.00553	0.89	1095.5	20.566	1096.7	27.3	1093.1	27.846			
FC-1	234	45271	1.52	1.94998	0.0577	0.18693	0.00422	0.76	1098.44	18.0726	1104.7	20.93	1086	35.126			
FC-1	332	177902	1.98	1.99366	0.0447	0.19061	0.0034	0.80	1113.36	13.7956	1124.68	16.744	1091.3	24.57			
FC-1	163	68111	1.85	1.99861	0.062	0.19113	0.0047	0.80	1115	19.11	1127.5	23.114	1090.8	33.852			
FC-1	248	31892	1.56	1.95095	0.0511	0.18805	0.0035	0.71	1098.77	15.9978	1110.8	17.29	1075	33.67			
FC-1	286	84360	1.72	1.96519	0.0542	0.1872	0.00393	0.76	1103.66	16.8896	1106.2	19.474	1098.7	32.578			
FC-1	478	101852	2.06	1.95253	0.0703	0.18633	0.00626	0.93	1099.3	22.022	1101.5	30.94	1095.1	24.024			
FC-1	180	69528	2.09	1.96703	0.0492	0.18624	0.0035	0.75	1104.29	15.3244	1100.97	17.3082	1110.8	30.03			
FC-1	548	135063	1.74	1.9303	0.0683	0.18236	0.00558	0.86	1091.6	21.476	1079.9	27.664	1115.2	32.76			
FC-1	295	106194	1.77	1.94821	0.0507	0.18452	0.00373	0.78	1097.83	15.8886	1091.6	18.382	1110.2	29.666			
FC-1	496	1828515	1.73	1.93792	0.0651	0.18606	0.00554	0.89	1094.3	20.384	1100	27.482	1082.9	28.028			
FC-1	1284	671896	1.84	1.93159	0.0761	0.18509	0.00655	0.90	1092.1	24.024	1094.7	32.396	1086.8	31.304			
FC-1	136	70296	2.21	1.95987	0.0564	0.18372	0.00408	0.77	1101.84	17.5994	1087.3	20.202	1103.7	33.306			
FC-1	452	61153	2.46	1.94648	0.067	0.18594	0.0055	0.86	1097.2	20.93	1099.3	27.118	1090.1	32.032			
FC-1	1111	666076	1.32	1.99996	0.0704	0.18926	0.00575	0.87	1115.5	21.658	1117.4	28.392	1111.9	31.486			
FC-1	233	1071825	1.61	1.95153	0.0492	0.18532	0.0035	0.75	1098.97	15.3972	1095.97	17.3264	1104.9	30.394			
R33	49	6239	1.52	0.51266	0.0252	0.06754	0.00142	0.43	420.24	15.3972	421.32	7.8078	414.3	90.272			
R33	45	52106	2.41	0.55665	0.0233	0.07188	0.00178	0.59	449.34	13.832	447.47	9.737	458.9	68.25			
R33	204	22312	1.27	0.51004	0.0138	0.06686	0.0013	0.70	418.48	8.4448	417.21	7.1526	425.5	39.312			
R33	184	22396	1.32	0.51274	0.0157	0.06618	0.00142	0.70	420.29	9.5914	413.1	7.8078	459.9	44.044			
R33	304	41050	1.15	0.51409	0.0155	0.0662	0.00146	0.73	421.2	9.464	413.22	8.0262	465.1	41.496			
R33	961	287889	1.53	0.50794	0.0194	0.06614	0.00216	0.85	417.06	11.8846	412.86	11.8846	440.4	40.768			
R33	84	9361	1.44	0.5148	0.0192	0.06678	0.00174	0.70	421.67	11.7026	416.73	9.5732	448.8	53.872			
R33	667	1186073	1.82	0.51598	0.0219	0.06762	0.00253	0.88	422.46	13.3406	421.8	13.9048	426.1	40.95			
R33	197	42202	1.21	0.51054	0.0147	0.06696	0.0012	0.64	418.81	8.9908	417.81	6.6066	424.3	44.954			
R33	149	789391	1.33	0.52358	0.0175	0.0681	0.00174	0.76	427.54	10.6106	424.7	9.555	442.9	44.044			
R33	200	54154	1.26	0.50823	0.0168	0.06541	0.00161	0.75	417.26	10.3012	408.44	8.8634	466.3	44.044			
R33	251	18866	1.12	0.50945	0.0199	0.06563	0.00165	0.65	418.08	12.1758	409.77	9.0818	464.2	59.878			
R33	94	35561	1.30	0.52117	0.0144	0.06881	0.0013	0.69	425.93	8.7542	428.98	7.1344	409.5	40.768			
R33	204	67829	1.26	0.52115	0.0181	0.06774	0.00183	0.78	425.92	10.9928	422.53	10.0464	444.3	44.044			
R33	215	25283	1.23	0.51526	0.0165	0.06743	0.00169	0.78	421.98	10.0646	420.65	9.282	429.2	40.586			
SL2	857	69909	6.13	0.72722	0.0241	0.08888	0.00272	0.92	554.92	12.8856	548.91	14.651	579.7	25.662			
SL2	385	31192	5.48	0.72586	0.0289	0.08967	0.00321	0.90	554.12	15.47	553.59	17.29	556.3	34.398			
SL2	394	169299	5.38	0.73928	0.0235	0.09055	0.00241	0.84	561.99	12.4852	558.79	12.9584	574.9	34.216			
SL2	109	23762	13.15	0.75535	0.0269	0.09076	0.00252	0.78	571.32	14.1596	560.03	13.559	616.5	43.862			
SL2	62	6358	9.15	0.74119	0.0283	0.0922	0.0021	0.59	563.1	15.015	568.54	11.284	541.2	61.334			
SL2	572	88857	6.24	0.71753	0.0255	0.0887	0.00268	0.85	549.21	13.7228	547.84	14.4326	554.9	37.128			
SL2	843	78553	6.18	0.72352	0.0243	0.08928	0.00264	0.88	552.74	13.0312	551.28	14.2142	558.8	31.668			
SL2	733	73164	5.79	0.72529	0.0225	0.08936	0.00247	0.89	553.78	12.0484	551.75	13.3042	562.2	28.028			
SL2	706	270556	5.86	0.71451	0.025	0.08811	0.00266	0.86	547.42	13.468	544.35	14.3416	560.2	35.49			
SL2	343	105312	1.18	0.77059	0.0205	0.09395	0.00188	0.75	580.1	10.7016	578.86	10.0828	585	34.762			
SL2	844	352706	6.13	0.72712	0.0227	0.0899	0.00236	0.84	554.86	12.1394	554.95	12.7036	554.5	33.67			
SL2	150	53742	10.41	0.73382	0.0223	0.09093	0.00216	0.79	558.79	11.8846	561.03	11.6116	549.7	36.946			
SL2	104	22060	13.24	0.73768	0.0243	0.09232	0.00209	0.68	561.05	12.922	569.24	11.2294	528	48.23			
SL2	371	96604	5.35	0.73337													

#### Appendix D4.1: Sample locations

sample #	Lat	Long	Description
49DTB19	47.05626	-116.43749	staurolite gt mu qf schist
50DTB19	46.99916	-116.35958	gt mica qf schist
51DTB19	46.98515	-116.31361	gt mu qf schist
59DTB19	47.02315	-115.46378	gt ky schist
52DTB19	47.01509	-115.81738	gt amphibolite

## Appendix D4.2: Bulk Rock Geochemistry

		49DTB19	50DTB19	51DTB19	59DTB19	52DTB19		
Wt. %	TiO2	79.9	0.617	0.572	0.597	0.686	1.888	
	Cr2O3	151.99	0.009	0.009	0.009	0.009	0.002	
	SiO2	60.09	71.19	67.63	61.21	62.1	58.01	
	Al2O3	101.96	15.02	17.68	21.47	16.57	13.53	
	CaO	56.08	0.28	0.27	0.21	0.39	6.64	
	MgO	40.31	1.73	1.72	1.59	3.49	3.24	
	FeO	71.85						
	MnO	70.94	0.041	0.04	0.04	0.114	0.228	
	Fe2O3	159.69	4.6	5.39	6.25	9.8	13.31	
	K2O	94.2	3.57	3.85	5.06	4.14	0.262	
	Na2O	61.98	0.89	0.69	0.89	0.57	2.53	
	H2O (LOI)	18.015	1.8	1.96	2.5	1.86	-0.04	
	Total		99.75	99.81	99.83	99.73	99.60	
	Moles	TiO2		0.008	0.007	0.007	0.009	0.024
		Cr2O3		0.000	0.000	0.000	0.000	0.000
SiO2			1.185	1.125	1.019	1.033	0.965	
Al2O3			0.147	0.173	0.211	0.163	0.133	
CaO			0.005	0.005	0.004	0.007	0.118	
MgO			0.043	0.043	0.039	0.087	0.080	
FeO			0.006	0.008	0.009	0.014	0.019	
MnO			0.001	0.001	0.001	0.002	0.003	
Fe2O3			0.012	0.014	0.016	0.025	0.033	
K2O			0.038	0.041	0.054	0.044	0.003	
Na2O			0.014	0.011	0.014	0.009	0.041	
H2O			0.100	0.109	0.139	0.103	-0.002	
total			1.558	1.536	1.512	1.494	1.417	
NF			64.171	65.109	66.153	66.922	70.574	
Normalized MOLAR %		TiO2		0.496	0.466	0.494	0.575	1.668
	Cr2O3		0.000	0.000	0.000	0.000	0.000	
	SiO2		76.024	73.279	67.386	69.161	68.131	
	Al2O3		9.453	11.290	13.930	10.876	9.365	
	CaO		0.320	0.313	0.248	0.465	8.356	
	MgO		2.754	2.778	2.609	5.794	5.673	
	FeO		0.411	0.488	0.575	0.913	1.307	
	MnO		0.037	0.037	0.037	0.108	0.227	
	Fe2O3		0.739	0.879	1.036	1.643	2.353	
	K2O		2.432	2.661	3.553	2.941	0.196	
	Na2O		0.921	0.725	0.950	0.615	2.881	
	H2O		6.412	7.084	9.180	6.910	-0.157	
	total		100.000	100.000	100.000	100.000	100.000	
	MnNCKFMASHTO	TiO2		0.50	0.47	0.49	0.57	1.67
		SiO2		76.02	73.28	67.39	69.16	68.13
Al2O3			9.45	11.29	13.93	10.88	9.37	
CaO			0.32	0.31	0.25	0.47	8.36	
MgO			2.75	2.78	2.61	5.79	5.67	
FeO			3.37	4.00	4.72	7.48	10.72	
MnO			0.04	0.04	0.04	0.11	0.23	
O			0.74	0.88	1.04	1.64	2.35	
K2O			2.43	2.66	3.55	2.94	0.20	
Na2O			0.92	0.72	0.95	0.62	2.88	
H2O			6.41	7.08	9.18	6.91	-0.16	
total			102.96	103.52	104.14	106.57	109.41	

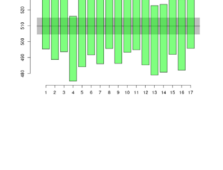
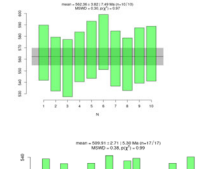
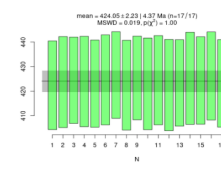






# Appendix D4.5: Monazite U/Pb Standards

Spot ID	238U conc.		232Th conc.		Isotopic Ratios					Apparent Ages - Calculated in Isoplot R (Ma)				Discord		error %	Concordia Age - Calculated in Isoplot R		Concordance within uncertainties?
	ppm		ppm		238U/206Pb	±2σ PRCP	207Pb/206Pb	±2σ PRCP	rho	238U/206Pb	±2σ PRCP	207Pb/206Pb	±2σ PRCP	[7/5-6]/[7/6+100]	Age		±2σ PRCP		
	ppm	ppm	ppm	ppm															
<b>MZ_Standard</b>																			
44050 - 1.d	3248.51804	24978.56992	14.7671698	0.6645920	0.05476356	0.00225887	0.16382081	422.39	9.2	401.6	46.2	-5.18	2.18%						
44050 - 2.d	3251.19098	24977.18427	14.7205493	0.6778399	0.05609394	0.00272337	0.15283659	423.69	9.44	455.1	34.1	6.90	2.23%						
44050 - 3.d	3256.09092	24979.75095	14.6643309	0.6520222	0.054959327	0.00272929	0.15979951	424.42	8.86	394.4	35.4	-7.62	2.11%						
44050 - 4.d	3250.18736	24974.83055	14.7118746	0.6739385	0.054984742	0.00220766	0.20976546	423.93	9.37	394.2	45.4	-7.54	2.21%						
44050 - 5.d	3254.04414	24983.83745	14.7420925	0.6556377	0.055755627	0.00283225	0.20202096	423.07	9.05	440.9	36.6	4.04	2.14%						
44050 - 6.d	3248.08667	24961.98752	14.6880347	0.6672523	0.056628978	0.001954879	0.23786607	424.59	9.33	452.5	38.7	6.17	2.20%						
44050 - 7.d	3253.32618	25016.10892	14.8370717	0.6292769	0.054922245	0.001749718	0.46378657	426.57	8.98	382.6	36.2	-11.49	2.11%						
44050 - 8.d	3253.32617	24966.10892	14.7657421	0.6729929	0.056406366	0.002412084	0.06338865	423.43	9.32	427.4	36.5	1.16	2.21%						
44050 - 9.d	3250.32884	24985.93038	14.6605939	0.6582894	0.056064741	0.002834888	0.45623385	425.36	8.68	453.9	36.3	6.29	2.04%						
44050 - 10.d	3253.170387	24973.48726	14.7462113	0.6632462	0.055153767	0.002005631	0.05575711	423.96	9.51	417.4	40.7	-3.33	2.25%						
44050 - 11.d	3250.487202	24977.77486	14.6931274	0.6621268	0.055642112	0.002070654	0.33869222	424.45	9.26	437.1	41.4	2.89	2.18%						
44050 - 12.d	3253.483473	24975.5121	14.7641995	0.6634083	0.054917708	0.002759353	0.10423664	423.47	9.48	406	35.9	-4.06	2.14%						
44050 - 13.d	3250.527148	24980.60205	14.7325161	0.6645546	0.056214515	0.001998813	0.29633513	423.35	8.96	436.4	40.4	2.99	2.12%						
44050 - 14.d	3251.529411	24978.71215	14.6626404	0.6634843	0.055161651	0.001947405	0.24097567	425.22	9.56	417.8	38.4	-1.78	2.23%						
44050 - 15.d	3250.624866	24972.69189	14.6970628	0.6694261	0.055146079	0.002848421	0.31921707	424.84	9.07	412.1	38.2	-3.74	2.14%						
44050 - 16.d	3251.079071	24978.82057	14.81380304	0.6486888	0.054626821	0.001927496	0.39671866	426.17	9.15	396	38.6	-7.62	2.15%						
44050 - 17.d	3250.973754	24974.27677	14.73986463	0.6532049	0.056089394	0.002802246	0.26016504	423.15	9.07	443.4	35.9	5.90	2.14%						
<b>MZ_Managony</b>																			
managony - 1.d	2445.312446	156050.9537	10.90078617	0.492967	0.056578176	0.002116638	0.35592706	565.8	12.2	474.1	41.4	-19.34	2.16%						
managony - 2.d	1146.417887	9321.71326	11.10473644	0.4897902	0.05884837	0.00229602	0.07440909	555.8	12	597	41.6	6.90	2.18%						
managony - 3.d	1113.705812	11370.2088	11.17970222	0.5366161	0.062124878	0.00294254	0.0910908	553.3	12.7	647.4	51.6	14.69	2.96%						
managony - 4.d	1308.763436	100317.2888	10.97415886	0.4498898	0.050474873	0.00221505	0.22781197	562.2	11	583.5	40.4	1.65	1.96%						
managony - 5.d	1029.294201	95103.9338	10.84893796	0.5507053	0.050349495	0.00233375	0.33783994	568.9	12.7	605.6	42.2	6.77	2.21%						
managony - 6.d	1538.781399	97030.14176	10.71686787	0.4754021	0.057623647	0.002289983	0.26911434	575.1	12.2	512.2	38.8	-12.28	2.12%						
managony - 7.d	1415.802273	118383.1438	11.00542994	0.5213989	0.050774697	0.00237869	0.3914022	566.6	12.2	629.6	41.7	10.96	2.18%						
managony - 8.d	1286.84136	115460.9165	11.10713846	0.4862068	0.04609311	0.002380442	0.17325293	557.7	11.7	551.1	40.6	0.83	2.11%						
managony - 9.d	1485.520272	98267.14296	10.95049834	0.4984727	0.058776721	0.002248044	0.03355068	563.3	12.3	557.9	41.7	-0.97	2.18%						
managony - 10.d	1460.030881	114848.716	10.91794175	0.4900237	0.050989357	0.002003949	0.26956143	564.9	12.2	451.1	40.3	-20.23	2.04%						
<b>MZ_Stern</b>																			
stern - 1.d	4736.046586	66468.49277	11.99881284	0.5357428	0.056965475	0.005837118	0.21543503	517.2	11.1	489.2	35.6	-5.72	2.15%						
stern - 2.d	2695.972093	45105.88994	12.14514313	0.5185263	0.05889079	0.002461033	0.30780262	510.1	10.9	486.2	35.6	-4.92	2.14%						
stern - 3.d	3076.606252	48353.88877	12.0280866	0.5232123	0.057149511	0.002848261	0.30794509	514.8	10.8	497.7	35.7	-3.44	2.10%						
stern - 4.d	4784.554834	58795.04526	12.51354096	0.5473657	0.059937215	0.001938613	0.35577088	495.6	10.4	579.8	36.5	14.52	2.20%						
stern - 5.d	2784.139366	5254.445454	12.2408373	0.5602668	0.057333485	0.00277814	0.3995708	506.2	11.3	503.4	34.1	-0.56	2.21%						
stern - 6.d	3922.771542	73201.66881	12.0483215	0.5160988	0.057977281	0.002847418	0.38769898	514	11.4	527.9	34.9	2.63	2.22%						
stern - 7.d	2784.139366	5254.445454	12.2080979	0.5471817	0.057769505	0.001321318	0.40486958	507.5	10.9	521.1	32.9	2.61	2.15%						
stern - 8.d	2276.051091	53168.36894	11.94280212	0.5359883	0.056848146	0.001742985	0.30533073	518.4	11.6	484.7	38.8	-6.95	2.24%						
stern - 9.d	3179.400178	40554.00829	12.2020957	0.5401742	0.056603985	0.002814348	0.34931286	507.8	11	476.5	35.6	-6.57	2.17%						
stern - 10.d	3460.870842	46554.85206	12.0201748	0.5394661	0.056904917	0.002762526	0.39373491	515.6	11.4	487.6	34.2	-3.74	2.12%						
stern - 11.d	4318.881506	58444.67977	11.9712849	0.546123	0.05813312	0.001940219	0.25752547	517.2	11.3	538	36.7	2.96	2.18%						
stern - 12.d	3073.227963	51315.0925	12.10964316	0.5764623	0.05741018	0.00193389	0.27096400	507.9	11.5	506.8	36.6	-0.22	2.20%						
stern - 13.d	4199.287748	64633.97244	12.37891218	0.5734718	0.064516	0.002816375	0.38748264	500.9	11.2	548.2	34.9	8.63	2.24%						
stern - 14.d	3079.228353	58119.96889	12.24412108	0.5287769	0.058512047	0.002004994	0.30838507	507.2	10.9	502.2	36	8.72	2.17%						
stern - 15.d	4027.241364	69071.79933	12.46509791	0.56072	0.05947495	0.00136623	0.2410713	514.1	11.3	506.2	33.9	-1.56	2.20%						
stern - 16.d	3463.582117	47747.9531	12.2813399	0.579812	0.067944284	0.001748046	0.02962402	504.5	11.5	480.6	33.9	-4.97	2.28%						
stern - 17.d	3163.348482	52047.80211	12.0512021	0.5454383	0.05830589	0.002709795	0.31818089	518	11.4	480.7	33.8	-20.96	2.20%						



# Appendix D4.6: Apatite U/Pb Results

Spot ID	238U conc.		232Th conc.		LA-ICP-MS Isotopic Ratios				
	ppm	±2σ INT	ppm	±2σ INT	238U/206Pb	±2σ PROP(%)	207Pb/206Pb	±2σ PROP(%)	rho
<b>49DTB19-AM</b>									
49DTB19-AM - 1.d	16.16910767	3.228772921	6.266060947	1.83720746	4.293569588	0.21663715	0.488298043	0.02127382	-0.167226571
49DTB19-AM - 3.d	0.028389451	0.037877737	0.002204098	0.011649757	0.058712115	0.011967308	0.874558684	0.042604669	0.006547696
49DTB19-AM - 4.d	35.1595877	7.146132933	17.5105983	7.157249245	4.950247987	0.224118394	0.412430588	0.015954654	-0.227660892
49DTB19-AM - 5.d	24.23391162	11.39616167	44.62092556	28.20789548	3.402707272	0.164162408	0.534505407	0.019654507	-0.236544666
49DTB19-AM - 6.d	24.86574051	5.69960188	17.43436269	3.14521581	4.161335316	0.179802944	0.48185622	0.017159251	0.380247864
49DTB19-AM - 7.d	3.313193842	1.310232909	0.535579031	0.421344579	0.812581735	0.048005138	0.10603114	0.033549295	0.173599725
49DTB19-AM - 8.d	41.5663909	5.399168233	62.71821622	10.36190696	4.953794319	0.207791613	0.405398136	0.01312726	0.206359263
49DTB19-AM - 9.d	30.87608627	5.824303489	17.3808957	5.087780567	4.838783016	0.205660979	0.447947609	0.014795148	0.337125431
49DTB19-AM - 10.d	17.34523631	12.84427607	30.8301296	20.66744251	3.264803664	0.187252048	0.572643098	0.021658588	-0.390443828
49DTB19-AM - 11.d	12.38677357	2.861355745	8.093189264	3.860958982	2.876571339	0.14481904	0.616686097	0.023248286	-0.111916185
49DTB19-AM - 16.d	4.422098035	4.958586231	0.789597818	1.482665242	1.192635159	0.101630077	0.742946516	0.032634135	-0.55514932
49DTB19-AM - 17.d	4.46554985	1.154423546	0.925480296	0.458292322	1.474176564	0.076262964	0.775376638	0.028424385	0.353871563
49DTB19-AM - 18.d	7.844615223	1.371272965	1.533876357	0.495189772	1.979072812	0.091506875	0.684435325	0.025028344	0.197917347
49DTB19-AM - 20.d	55.40247874	3.368806374	26.59298814	2.167576033	6.488591998	0.276501034	0.31068806	0.011447432	0.182385994
49DTB19-AM - 23.d	16.96791556	4.212323846	17.03223669	4.245602353	4.962128674	0.314781042	0.386465924	0.029054659	0.376476478
49DTB19-AM - 26.d	39.30371567	1.954011549	11.58922897	1.461058333	5.898956582	0.321502697	0.402999507	0.020328634	0.633787718
49DTB19-AM - 27.d	7.88623735	0.995867764	11.59720547	2.403037098	2.1919273	0.105474497	0.700490514	0.026869957	0.417499922
49DTB19-AM - 28.d	6.667566598	1.013370613	10.31899021	7.983077081	1.832010397	0.080841571	0.710425273	0.02441106	0.454407317
49DTB19-AM - 29.d	2.660113848	1.142757859	0.985311425	0.684239342	1.09711545	0.052890202	0.783075506	0.031129599	0.282704629
49DTB19-AM - 34.d	40.40172903	26.76539095	34.61111615	18.09300045	5.329109297	0.274842196	0.37270256	0.015046028	-0.538995655
49DTB19-AM - 35.d	97.54443104	27.232530835	89.90986547	17.28400687	7.531196467	0.311733765	0.255358325	0.009754442	0.012412085
49DTB19-AM - 37.d	2.320421023	0.495235113	0.36166923	0.38560728	0.738672212	0.037259502	0.767675041	0.028653981	0.237336767
49DTB19-AM - 38.d	5.2036824	1.615117416	9.566973263	3.875009661	1.405846019	0.071773112	0.741778976	0.026348777	0.14050001
49DTB19-AM - 39.d	4.134693337	1.69243573	0.435014242	0.251418593	1.00631578	0.08282848	0.76900225	0.037557042	0.224159471
49DTB19-AM - 40.d	3.977711821	0.970530998	1.669056205	0.973667679	1.579743305	0.087958993	0.709948834	0.042155589	0.363020091
49DTB19-AM - 42.d	2.759235582	0.569117949	0.40528838	0.196136214	0.884217743	0.041836714	0.797525282	0.029914831	0.404813914
49DTB19-AM - 43.d	3.024803457	1.10707385	0.475645798	0.332707673	0.899636334	0.043958646	0.778552831	0.029252909	0.134752933
49DTB19-AM - 44.d	2.189912805	0.331930743	0.202973789	0.117415242	0.710752047	0.032202278	0.825261038	0.029125107	0.466622987
49DTB19-AM - 45.d	0.75947272	1.53187956	0.192473296	0.133839369	0.358722727	0.033403995	0.877364134	0.033199385	-0.023345483
49DTB19-AM - 47.d	8.002808886	3.317626745	2.76381119	1.539251774	2.441126389	0.144055525	0.716564917	0.027652821	-0.192531677
49DTB19-AM - 48.d	8.009797223	2.306173947	2.192078268	0.875410229	2.157780902	0.099991452	0.684006905	0.025348245	0.382758257
49DTB19-AM - 49.d	2.21127492	0.490144441	0.188606164	0.112240316	0.708572015	0.033293332	0.813088479	0.02876897	0.458084892
49DTB19-AM - 51.d	5.650497929	3.531511664	2.286983497	1.879816372	1.618933328	0.096297014	0.710293608	0.026313466	-0.190388613
49DTB19-AM - 54.d	2.138071423	0.846693215	0.595140353	0.389333385	0.933605352	0.049256498	0.832658355	0.031603976	0.225865775
49DTB19-AM - 55.d	0.16250941	0.410881777	0.053395607	0.092214278	0.113915187	0.012375797	0.865993985	0.033137099	-0.114124128
49DTB19-AM - 57.d	4.596504364	1.071078561	0.41573973	0.158629467	1.185474173	0.054529918	0.758515044	0.027000290	0.379857184
49DTB19-AM - 59.d	1.448022703	1.190721116	0.17192939	0.146946523	0.74076446	0.148567505	0.789565486	0.025709707	0.252115133
49DTB19-AM - 60.d	2.852929225	1.576529035	1.563867619	1.309603077	0.696500915	0.043833845	0.826239706	0.028438251	-0.031278173
<b>51DTB19-AM</b>									
51DTB19-AM - 1.d	11.63337735	1.838209317	0.181500918	0.122528037	12.0360767	0.593185029	0.485117918	0.028135488	0.628914966
51DTB19-AM - 2.d	11.08094902	1.519430046	0.166653133	0.112339602	10.75361512	0.55723492	0.484503926	0.026266115	0.537198453
51DTB19-AM - 3.d	38.66091604	25.27232851	1.84309789	1.64265776	3.179233331	2.041284779	0.377171922	0.022447156	-0.290348763
51DTB19-AM - 4.d	12.6407077	1.468906989	0.186399311	0.121181734	12.79876451	0.63722049	0.473141103	0.027402197	0.666565283
51DTB19-AM - 5.d	12.38942254	1.79002887	0.188311922	0.136200604	12.40007895	0.665336755	0.448851572	0.026533946	0.648474781
51DTB19-AM - 6.d	26.38224325	8.112203163	1.14209189	0.627447792	20.31769523	1.07999912	0.398411968	0.023442412	0.609023037
51DTB19-AM - 8.d	12.39906707	1.947046842	0.203632665	0.129524166	13.45942058	0.669176784	0.488435153	0.021863031	0.643008786
51DTB19-AM - 9.d	18.38178909	2.61017016	2.283947606	2.286607828	7.883203441	0.418303792	0.33320451	0.018003740	0.531279642
51DTB19-AM - 10.d	14.20209589	9.539038898	0.543876731	0.888421203	8.63300039	0.543252475	0.342459628	0.027423568	0.483829799
51DTB19-AM - 11.d	53.0738374	12.68973276	3.000203205	0.89146889	37.57886428	1.901888	0.349380606	0.027186246	0.573507752
51DTB19-AM - 12.d	14.32441191	1.801157289	0.13722737	0.142087392	12.97940644	0.654577166	0.449757166	0.030135283	0.628143022
51DTB19-AM - 13.d	1.711416455	0.774403747	0.170138435	0.134644359	2.22145169	0.164614246	0.579811199	0.04709858	0.123483308
51DTB19-AM - 14.d	14.41942448	4.017609207	0.27769654	0.152655781	18.85666583	0.686315639	0.383897829	0.023090017	0.522138555
51DTB19-AM - 15.d	13.19935902	2.196970995	0.332629224	0.166748703	8.97825713	0.434530803	0.390441418	0.019247796	0.201013219
51DTB19-AM - 16.d	12.6866107	1.648231646	0.246843115	0.12818073	14.17486392	0.738086297	0.520077184	0.029089544	0.686642329
51DTB19-AM - 17.d	28.8411892	6.801183734	1.260383354	0.49938882	23.69730279	1.321194486	0.354471345	0.023700553	0.541465246
51DTB19-AM - 18.d	6.828717745	0.89552495	0.141315931	0.114727492	6.355256257	0.5156394	0.368161124	0.039177434	0.654620991
51DTB19-AM - 19.d	15.96177443	2.670647669	0.286778032	0.148782558	8.95948174	0.402016546	0.255308401	0.012967921	0.415605779
51DTB19-AM - 20.d	8.081757339	1.806233991	0.271490946	0.140842601	8.079519726	0.429005412	0.534254435	0.041103288	0.481413228
51DTB19-AM - 21.d	17.17863222	4.708266543	0.324602344	0.230763209	18.08338159	1.001917432	0.429315223	0.024505894	0.482138384
51DTB19-AM - 24.d	16.79203442	3.363584913	0.79767868	0.847637425	15.93427279	0.91476153	0.487302844	0.030709424	0.572430327
51DTB19-AM - 25.d	18.50754991	13.62317765	0.660555221	0.292218897	16.57044916	1.134779324	0.427088332	0.026877877	0.115155444
51DTB19-AM - 26.d	10.22657271	7.291678444	0.545941639	0.35214014	11.16766457	0.839485714	0.393858643	0.038657859	-0.151090045
51DTB19-AM - 28.d	16.33664943	6.905920978	0.404554824	0.479740806	16.38326173	0.993346494	0.422167628	0.029920958	0.4824416
51DTB19-AM - 29.d	13.74598407	1.21326282	0.19822327	0.109510322	12.75438979	0.670081852	0.449218785	0.030758927	0.681163106
51DTB19-AM - 31.d	16.62415615	4.755789044	0.806267775	0.854156591	10.40401002	0.557560817	0.40469445	0.036437574	0.170377467
51DTB19-AM - 33.d	13.72950283	2.140597483	0.217766906	0.132032192	13.02237063	0.656293837	0.454686058	0.0243631	0.356879079
51DTB19-AM - 34.d	36.05165866	11.67145678	1.692652205	0.601028385	30.94603359	1.798763087	0.326368788	0.0286367	0.127197221
51DTB19-AM - 36.d	15.03464115	1.67236717	0.261657893	0.137189027	11.66321496	0.551144547	0.363930142	0.019253875	0.455978741
51DTB19-AM - 37.d	1.26500302	1.385852664	0.021722665	0.0					

# Appendix D4.6 continued.

Spot ID	238U conc.		232Th conc.		LA-ICP-MS Isotopic Ratios				
	ppm	±2σ INT	ppm	±2σ INT	Z38U/206Pb	±2σ PROP(2%)	207Pb/206Pb	±2σ PROP(2%)	rho
<b>52DTB19-AM</b>									
52DTB19-AM - 1.d	9.2334247	0.877672555	0.792702702	0.271824098	28.56222961	1.891494129	0.372533945	0.033636146	0.487984445
52DTB19-AM - 2.d	9.994556318	1.918500267	0.51598314	0.13552251	30.14502133	1.83809213	0.38127465	0.03412088	0.593188828
52DTB19-AM - 3.d	13.93637541	1.015543538	0.673978474	0.23463708	41.15623592	2.418169924	0.34526311	0.028519859	0.443112367
52DTB19-AM - 4.d	12.35049477	0.202687834	1.002805303	0.281178973	37.10959604	2.27432351	0.370636039	0.03484441	0.708552934
52DTB19-AM - 5.d	12.3143219	1.017884323	0.616436866	0.188158581	38.6534223	2.372692547	0.351526204	0.032998715	0.60962933
52DTB19-AM - 6.d	15.08364696	4.44601778	0.590215746	0.241874836	36.18714115	4.382742157	0.29137007	0.03451983	0.54818956
52DTB19-AM - 7.d	12.10759867	1.848668388	0.59590315	0.167905257	33.94833266	2.042092514	0.371998128	0.033041752	0.666770984
52DTB19-AM - 8.d	14.86976064	1.902687206	0.74635972	0.289895124	43.97959007	0.311339792	0.311339792	0.029839396	0.578663996
52DTB19-AM - 9.d	10.61557295	1.567492737	0.604603355	0.18413502	31.54609219	1.868025865	0.377399068	0.035969574	0.511051174
52DTB19-AM - 10.d	14.77397792	2.6329433	0.77531378	0.22618509	52.57551215	3.284210254	0.312842399	0.028295198	0.4820629
52DTB19-AM - 11.d	11.38246415	1.062490536	0.666640384	0.207821439	31.34534789	1.891810413	0.355057211	0.030325233	0.61021669
52DTB19-AM - 12.d	11.78630647	1.898152857	0.732657119	0.242117965	32.31999336	1.966385386	0.379015483	0.02996535	0.586915998
52DTB19-AM - 13.d	10.33095822	1.877235825	0.503698151	0.216191683	29.75042128	1.789409667	0.396884655	0.033239531	0.499715699
52DTB19-AM - 14.d	10.53886729	1.038214256	0.50054518	0.192279581	31.54048077	1.815306544	0.409267632	0.034414626	0.615039539
52DTB19-AM - 15.d	9.765312854	1.671562041	1.190489416	0.403789706	40.56566355	3.05571659	0.304171262	0.02904018	0.333522128
52DTB19-AM - 16.d	22.71020434	3.823799755	1.153081681	0.429563135	67.27967822	4.15251096	0.250761103	0.024919808	0.408741689
52DTB19-AM - 17.d	21.6160379	3.294941681	1.967737097	1.130919197	75.72675294	6.724085072	0.201252551	0.024479056	0.532627080
52DTB19-AM - 19.d	10.42081702	2.358135627	0.676253874	0.254850284	31.35578875	1.930267556	0.402219126	0.034251164	0.538273947
52DTB19-AM - 20.d	3.250940763	2.165263618	1.257872147	1.159142558	23.61923107	2.918436487	0.376046511	0.037556774	0.423140235
52DTB19-AM - 21.d	11.81324946	1.77388635	0.551768635	0.169724769	37.63949879	2.352847604	0.37490476	0.031741106	0.615039539
52DTB19-AM - 22.d	15.05765212	1.071652071	1.390489416	0.403789706	40.56566355	3.05571659	0.304171262	0.02904018	0.333522128
52DTB19-AM - 23.d	2.037003353	0.91322981	0.243866534	0.407286397	22.99172573	4.477221573	0.467679578	0.10771887	0.683100848
52DTB19-AM - 24.d	13.45639553	2.515449506	0.853053305	0.651174534	38.85321525	2.347498756	0.336393936	0.025842321	0.306472868
52DTB19-AM - 25.d	9.996475126	1.019726733	0.559204508	0.206412176	31.38682224	1.848496085	0.341739661	0.024139606	0.480937457
52DTB19-AM - 26.d	16.97091987	2.252321129	0.353157794	0.186910535	72.90228271	5.29615969	0.257268884	0.02929772	0.463712297
52DTB19-AM - 27.d	15.09203666	2.212345743	0.491982555	0.205491257	59.9515164	4.050876904	0.2818305	0.036481756	0.70518795
52DTB19-AM - 28.d	16.11843802	2.036075843	0.611740273	0.175159881	44.4459764	4.977305361	0.283319423	0.038478013	0.549941233
52DTB19-AM - 30.d	6.873751765	4.32970566	0.55110211	0.42582686	24.90241181	2.812278332	0.461450571	0.049325063	0.028137587
52DTB19-AM - 31.d	13.2987512	2.834915179	0.921643136	0.498893678	37.21654491	2.173865767	0.363560062	0.030835199	0.563116607
52DTB19-AM - 32.d	15.99974618	3.937639858	0.559892131	0.26298096	69.42649232	4.783253255	0.229616484	0.029853409	0.437737465
52DTB19-AM - 33.d	5.31283197	2.173215973	0.936256795	0.828975071	8.778095656	0.923337478	0.439548336	0.039872438	0.237761018
52DTB19-AM - 34.d	14.22112055	1.60076662	1.691191302	0.324619139	41.71554105	2.57813774	0.321389773	0.031296809	0.630468297
52DTB19-AM - 35.d	14.65550585	1.990457137	1.751559881	0.383032845	44.4459764	2.720113768	0.306967607	0.027596466	0.551919328
52DTB19-AM - 36.d	12.92414496	2.074602234	1.246647184	0.235069505	36.21619115	2.225823725	0.357404161	0.029971125	0.402662367
52DTB19-AM - 37.d	13.58231248	4.708640314	1.457616955	0.58175463	36.24408202	2.664398424	0.40376635	0.036916312	0.294656687
52DTB19-AM - 38.d	16.50824557	2.580938478	1.860994512	0.516558889	40.51516154	2.437109138	0.356102723	0.029073919	0.426034051
52DTB19-AM - 39.d	11.43173784	1.515963045	0.621018802	0.237231398	36.47648174	2.234239165	0.367933409	0.033830603	0.424044853
52DTB19-AM - 41.d	9.769051414	0.85063816	0.928468733	0.261901792	40.42978373	3.438982894	0.322718438	0.031296809	0.451941183
52DTB19-AM - 43.d	13.94849741	4.981209646	1.386459853	0.255217808	44.78602816	3.80204144	0.295864326	0.031610112	0.41101112
52DTB19-AM - 44.d	12.29070431	2.043867717	3.371491702	1.155508795	55.8169546	1.521902717	0.369572496	0.032044114	0.23753612
52DTB19-AM - 45.d	11.16760747	0.973639614	0.45354892	0.163000519	35.48652872	2.184742308	0.363018273	0.032297842	0.524240205
52DTB19-AM - 46.d	19.57774547	1.254771835	1.624554614	0.544969704	55.37813025	3.20886411	0.270192151	0.024593983	0.240811828
52DTB19-AM - 47.d	10.21794226	2.157953091	0.441090124	0.195666675	37.70317874	1.620903615	0.332875216	0.040867343	0.362896294
52DTB19-AM - 48.d	14.25186057	1.535510315	0.888762798	0.308692986	51.26960286	3.528139759	0.288196346	0.028436356	0.604940886
52DTB19-AM - 49.d	17.79443808	1.072655512	1.072655512	0.270691373	36.49392928	0.75507998	0.344140426	0.030161681	0.340148808
52DTB19-AM - 50.d	15.54046019	4.846204047	1.015891329	0.396568866	45.89161013	3.40351671	0.368465115	0.041168881	0.108707805
52DTB19-AM - 51.d	17.65688662	2.641501693	1.23080348	0.292054067	52.35879463	3.214666891	0.293004866	0.026082925	0.503653475
52DTB19-AM - 52.d	8.592927503	1.049887857	0.506676051	0.224783178	28.46390753	2.569757639	0.372487506	0.045663041	0.48310271
52DTB19-AM - 53.d	11.70700657	3.780244211	0.834429483	0.342509768	29.56566617	2.447415592	0.404927849	0.047199424	0.394574072
52DTB19-AM - 54.d	17.77184571	1.936083787	1.09258078	0.32943298	53.82578677	3.439884359	0.311383173	0.03157276	0.675388326
52DTB19-AM - 55.d	17.03414982	1.153987777	0.710411775	0.26186977	70.96164164	4.056771103	0.20052828	0.020809209	0.467750322
52DTB19-AM - 56.d	10.57831113	1.970762569	0.556808556	0.214621956	40.43764122	2.57708449	0.349661356	0.048858907	0.488585907
52DTB19-AM - 57.d	15.67661536	3.01112461	0.959280112	0.29977541	49.57589233	3.07131589	0.301964729	0.027310527	0.358112133
52DTB19-AM - 58.d	12.85315122	2.451671413	1.009589341	0.33805063	40.08013812	2.465532183	0.330694489	0.027850703	0.59367157
52DTB19-AM - 59.d	14.7368052	1.155568703	0.321494547	0.163000519	45.6030354	2.896924236	0.319911126	0.030920485	0.631322852
52DTB19-AM - 60.d	14.15638012	2.174116663	1.072508571	0.315390133	43.4860699	2.768737341	0.342919667	0.03211931	0.605530153
52DTB19-AM - 61.d	19.83750159	2.614094737	1.166120312	0.305894245	59.53545391	4.551549412	0.296425125	0.034220676	0.475884254
52DTB19-AM - 62.d	13.39706582	3.69522232	0.520052126	0.248516197	39.07629306	2.269167489	0.318871961	0.02944631	0.483811786
52DTB19-AM - 63.d	15.9313156	2.137048389	0.853633032	0.283253361	31.704549	1.31522397	0.330841886	0.030841886	0.587715327
52DTB19-AM - 64.d	17.4124595	9.861731999	1.15192453	0.72076028	52.14527672	3.418028573	0.288818256	0.02689454	0.186193441
52DTB19-AM - 65.d	11.80242006	2.115277572	0.636906914	0.225748425	34.62721491	2.114724648	0.383104117	0.032513403	0.531057318
52DTB19-AM - 66.d	12.22406094	2.260556534	0.95706091	0.389496461	37.08876818	2.356237906	0.34166008	0.03098866	0.438915039
52DTB19-AM - 67.d	10.04370437	1.913446427	0.33837073	0.182373348	31.0194734	2.009651274	0.384283542	0.033178821	0.491394633
52DTB19-AM - 68.d	12.31201767	1.700886071	0.60934692	0.199932621	35.06126885	2.224089507	0.408504633	0.034374246	0.691679399
52DTB19-AM - 69.d	12.5638809	1.479453816	1.063066917	0.320349338	35.58955114	2.090402937	0.371929742	0.031183047	0.46595987
52DTB19-AM - 70.d	5.827220574	1.224027236	1.382032073	0.465485147	17.66395105	1.056747008	0.465025761	0.038692049	0.618481781
52DTB19-AM - 71.d	7.934749121	1.334500585	0.901886987	0.340394194	34.00394194	2.430178665	0.348048046	0.03823278	0.218868956
52DTB19-AM - 72.d	10.64089223	2.757628826	0.500351178	0.253993507	39.47245931	6.052836197	0.406509956	0.026820599	-0.21

# Appendix D4.6 continued.

Spot ID	238U conc.		232Th conc.		LA-ICP-MS Isotopic Ratios				
	ppm	±2σ INT	ppm	±2σ INT	238U/206Pb	±2σ PROP(2%)	207Pb/206Pb	±2σ PROP(2%)	rho
<b>590TB19-AM</b>									
590TB19-AM - 3.d	2.100379556	1.059529902	0.6669529051	1.941849255	0.573109155	0.031526071	0.846897153	0.037097102	0.342477266
590TB19-AM - 5.d	47.55065696	23.83707048	16.17172008	10.89673197	11.10887785	0.577823247	0.761665099	0.027558245	0.262546465
590TB19-AM - 6.d	15.60484452	2.871855487	3.128329292	0.883044859	3.69439325	0.197600528	0.812979557	0.038330247	0.395515931
590TB19-AM - 7.d	14.94387542	1.958715543	4.579852699	1.091711434	4.51954255	0.218487203	0.798803624	0.030008436	0.488083952
590TB19-AM - 9.d	9.02090032	4.305511732	3.165728834	0.71312707	1.783958999	0.109571796	0.788350605	0.02775008	-0.089138318
590TB19-AM - 11.d	6.666859786	0.697130932	0.665011951	0.251452377	1.778215063	0.080232357	0.82656627	0.028753458	0.267134285
590TB19-AM - 12.d	3.830627176	0.631219296	0.584815702	0.460232909	1.002414639	0.044364812	0.820260009	0.028713714	0.488900482
590TB19-AM - 13.d	2.546568497	0.595578444	0.380975309	0.17852553	0.513058147	0.024316728	0.864871706	0.028517143	0.277802176
590TB19-AM - 14.d	78.29558571	10.14215199	19.2574845	3.37892122	19.3289389	1.094774562	0.689931668	0.020110442	-0.050623436
590TB19-AM - 15.d	12.83630764	14.34720823	1.982765261	2.575798053	3.603793216	0.438060508	0.797624094	0.030178089	-0.041251072
590TB19-AM - 16.d	46.03631377	7.3993884575	9.992639294	1.741484201	9.743742614	0.448662398	0.752425001	0.027315773	0.212548216
590TB19-AM - 17.d	4.298995316	0.553136325	0.254609086	0.140991066	1.049594983	0.047766427	0.833266643	0.028777177	0.549898747
590TB19-AM - 18.d	0.511552	0.844226246	0.709297832	1.091796997	0.514400825	0.03913179	0.804013373	0.041280477	0.027553655
590TB19-AM - 19.d	17.38700627	20.30129281	3.627053292	4.746371704	4.457879038	0.540470477	0.805392214	0.02930823	-0.194561471
590TB19-AM - 20.d	4.464957326	0.54863078	1.187109923	1.241259423	1.150912866	0.051605481	0.82090169	0.028526637	0.495618319
590TB19-AM - 21.d	4.397947328	0.97965997	1.340389922	0.581635337	1.076360913	0.049800528	0.837624853	0.02804705	0.603145481
590TB19-AM - 22.d	4.041185783	0.586864661	0.293364875	0.153113877	1.143540304	0.05183013	0.82647137	0.029793766	0.534151675
590TB19-AM - 23.d	0.25005604	0.287652444	0.60658225	0.601348799	0.185487016	0.015998377	0.825148301	0.030709649	0.280115199
590TB19-AM - 24.d	0.140881417	0.115126337	0.308763203	0.14714698	0.14714698	0.051003792	0.867683646	0.03528088	0.306022853
590TB19-AM - 25.d	8.270320497	9.511588758	2.757916272	3.23817105	1.580628785	0.170713893	0.811203757	0.028120355	-0.055955419
590TB19-AM - 26.d	4.70004616	1.110607663	0.927329257	1.209617312	1.256732556	0.05806938	0.81521099	0.028118561	0.514076795
590TB19-AM - 27.d	3.920328354	2.410472353	2.031855613	1.311597216	0.781287527	0.052448079	0.84226157	0.027716173	0.122227409
590TB19-AM - 28.d	0.131286612	0.086395366	0.380625364	0.298682672	0.489951563	0.04494522	0.83489486	0.045139994	-0.053743422
590TB19-AM - 29.d	4.196779027	0.931006125	1.482025228	0.403879699	0.948003064	0.044999791	0.81860259	0.027394478	0.385556599
590TB19-AM - 30.d	0.25235408	0.298770623	1.447516222	2.439894533	0.425886064	0.035772299	0.802432189	0.038149568	0.311476695
590TB19-AM - 31.d	3.220871208	0.436069104	0.42730885	0.193284591	1.228116646	0.062958494	0.844664426	0.038329103	0.883763642
590TB19-AM - 32.d	38.91848368	8.4966695	9.42038672	1.807587974	12.1015058	0.560267077	0.750255768	0.025278669	0.477637411
590TB19-AM - 33.d	5.091146474	0.980175184	0.94466319	0.221252539	1.272780999	0.058271157	0.846226006	0.029377784	0.557919349
590TB19-AM - 34.d	0.137691971	0.095418681	0.336704008	0.1532724	0.475312623	0.040284676	0.827518569	0.043025197	0.069807996
590TB19-AM - 35.d	0.146840053	0.097319341	0.376516113	0.207934338	0.485401705	0.0411005	0.860406536	0.043120096	0.088647212
590TB19-AM - 36.d	4.24884379	0.897597861	0.300188462	0.156315654	1.235481119	0.05471758	0.847876471	0.036070716	0.476571727
590TB19-AM - 37.d	12.68450787	3.595076342	2.669431605	0.866623493	3.164958866	0.157495997	0.811776436	0.029643987	0.292118023
590TB19-AM - 38.d	0.180650755	0.124741281	0.727739599	0.203617472	0.985013968	0.171815867	0.830495729	0.048378707	0.131286998
590TB19-AM - 39.d	13.28471129	13.64540071	2.191551559	3.025545293	3.462574263	0.399427293	0.865274263	0.024045393	0.041045393
590TB19-AM - 40.d	21.28075165	13.57485238	3.810550105	2.718002706	0.075429801	0.352033631	0.806111987	0.027506691	0.189254295
590TB19-AM - 41.d	0.912220246	0.427250448	0.404320211	0.203466841	0.647893157	0.041937195	0.820481045	0.042801204	0.330419458
590TB19-AM - 42.d	4.443705769	0.984843463	0.308603998	0.155502246	1.224172726	0.054959129	0.836579925	0.029521075	0.461686424
590TB19-AM - 43.d	4.187447275	0.839757386	0.48005942	0.196109731	1.143278958	0.052232417	0.843791992	0.032175071	0.713764257
590TB19-AM - 44.d	16.70765496	3.04767196	4.780379364	1.067062015	3.320918808	0.149182921	0.761976026	0.048173807	0.267852293
590TB19-AM - 45.d	8.346490927	6.246499192	0.70751448	0.369796444	2.425456841	0.127309864	0.827374478	0.028112646	0.486636359
590TB19-AM - 46.d	3.40679711	1.870485595	0.664693028	0.93837602	0.828351697	0.055126988	0.837467163	0.029006884	0.073018256
590TB19-AM - 47.d	5.46815954	22.17585773	7.96441177	4.429716497	10.0468999	0.713572496	0.778986717	0.027346663	-0.116631294
590TB19-AM - 48.d	22.5663204782	1.529699448	0.710044527	0.145380398	0.086736729	0.802419474	0.802419474	0.029746286	-0.216681604
590TB19-AM - 49.d	1.824342521	0.682771157	13.77197937	7.47092854	0.333785088	0.016606679	0.749885544	0.025799964	0.151806811
590TB19-AM - 50.d	16.02916806	1.857272506	2.109961097	0.429272258	1.83702986	0.173087067	0.825514581	0.059722744	0.19817656
590TB19-AM - 51.d	4.152365259	0.82383252	0.70751448	0.369796444	1.100203443	0.048073681	0.837252807	0.029806759	0.456636359
590TB19-AM - 52.d	3.40679711	1.870485595	0.664693028	0.93837602	0.828351697	0.055126988	0.837467163	0.029006884	0.073018256
590TB19-AM - 53.d	5.46815954	22.17585773	7.96441177	4.429716497	10.0468999	0.713572496	0.778986717	0.027346663	-0.116631294
590TB19-AM - 54.d	22.5663204782	1.529699448	0.710044527	0.145380398	0.086736729	0.802419474	0.802419474	0.029746286	-0.216681604
590TB19-AM - 55.d	1.824342521	0.682771157	13.77197937	7.47092854	0.333785088	0.016606679	0.749885544	0.025799964	0.151806811
590TB19-AM - 56.d	16.02916806	1.857272506	2.109961097	0.429272258	1.83702986	0.173087067	0.825514581	0.059722744	0.19817656
590TB19-AM - 57.d	4.152365259	0.82383252	0.70751448	0.369796444	1.100203443	0.048073681	0.837252807	0.029806759	0.456636359
590TB19-AM - 58.d	9.041586807	2.788138642	1.086815875	0.71010919	2.639250886	0.15323972	0.813825368	0.029644498	0.075487651
590TB19-AM - 59.d	5.931587385	1.063698181	3.739986639	0.37036028	2.397326039	0.111538626	0.815859253	0.030831305	0.450063581
590TB19-AM - 60.d	0.180122057	0.178747804	0.875109757	0.93564961	0.473729912	0.038029883	0.856614913	0.042359269	0.207910285
590TB19-AM - 61.d	22.45696436	4.64395039	5.140344682	0.974867758	5.429484402	0.295021973	0.804291602	0.02775778	0.211650786
590TB19-AM - 62.d	3.132496615	1.001669717	1.063413207	0.454104891	0.961694909	0.047723435	0.825496677	0.030613505	0.458088656
590TB19-AM - 63.d	7.252020959	0.985259747	2.058456985	0.57862866	2.226021242	0.147411555	0.835883441	0.052183323	0.25215069
590TB19-AM - 64.d	13.42213805	6.742080065	7.820794347	1.707079138	3.931300008	0.19540635	0.802778309	0.027023535	0.246808798
590TB19-AM - 65.d	1.530799876	0.711948921	0.709147829	0.347083513	0.633211985	0.034737188	0.836247647	0.032091257	0.289906286
590TB19-AM - 66.d	0.129353205	0.0731417	0.347305765	0.25203359	0.336720426	0.027655718	0.836316963	0.032748555	0.261286855
590TB19-AM - 67.d	10.20393032	13.30430846	1.51312533	1.52699947	2.480571777	0.277187636	0.815385857	0.028449551	-0.017900225
590TB19-AM - 68.d	7.05764891	0.777176617	1.136997155	0.503016119	1.854814708	0.085042852	0.839645363	0.031164091	0.498295312
590TB19-AM - 69.d	21.0067211	2.822012704	7.141012825	2.333753474	5.288156731	0.244434767	0.789301415	0.028710603	0.639203005
590TB19-AM - 70.d	8.262478658	0.921414748	3.40385676	1.589953322	1.800278381	0.084128119	0.836752956	0.029637819	0.377866681
590TB19-AM - 71.d	4.567096499	1.171361742	0.692787161	0.820825015	1.304007109	0.059286533	0.832726134	0.030947373	0.559769343
590TB19-AM - 72.d	3.766132115	0.50797853	1.600273859	0.550885634	0.733018977	0.035700471	0.829441394	0.033827167	0.534867446
590TB19-AM - 73.d	0.480762999	0.441258725	2.09478559	1.498532165	0.092844633	0.010101404	0.827599981		



## Appendix D5.1: Sample Locations

**TABLE S1. LOCATIONS AND DESIGNATIONS FOR SAMPLES ANALYZED IN THIS PAPER**

Sample number	Geologic Unit	Location		Elevation (m asl)	Analysis	Location Description
		Latitude (°N)	Longitude (°W)			
<b>BH1</b>	Buffalo Hump	48.1952	117.91376	1335	LA-SS-ICPMS	Ridge E/SE. of Klein Meadows. Upper Ybq
<b>03DTB18</b>	Buffalo Hump	48.18471	117.93345	1323	LA-SS-ICPMS	Lower Ybq near Double eagle Quarry
<b>04DTB18</b>	Buffalo Hump	48.18413	117.93176	1320	LA-SS-ICPMS	Along road cut N of Double eagle Quarry. Mid-lower Ybq
<b>05DTB18</b>	Buffalo Hump	48.20175	117.9069	1070	LA-SS-ICPMS	Duplication of Ross 1992 locality
<b>34DTB19</b>	Buffalo Hump conglomerate individual cobble sampled	48.13122	117.99775	1295	LA-SS-ICPMS	Conglomerate bed at 55.5-57m on measured section 1

Note: Locations are based on datum WGS84.



# Appendix D5.2: U/Pb Results

Sample ID	Analysis date-time	Spot ID <sup>1</sup>	238U conc.			235U conc.			Isotopic Ratios			Assessment Ages - Calculated in August 8, 2014				Diseq <sup>2</sup>		Preferred age (50%)		
			ppm	±1σ	2σ	ppm	±1σ	2σ	238U/235U	235U/238U	207Pb/235U	206Pb/238U	207Pb/238U	206Pb/238U	207Pb/238U	206Pb/238U	(78-6/98)7e+100	Ma conf.	Ma	2-σ
BH1	2020-07-30 16:07:02.000 BH1-1540-1-Ld		306	7	340	7	8.02557	0.23716	0.68389	0.004110	0.118118	757	21	734	136			1.11	757	21
BH1	2020-07-30 16:08:48.000 BH1-1546-1-Ld		334	6	377	7	8.01230	0.23709	0.004096	0.004132	0.121823	758	21	734	135			0.70	758	21
BH1	2020-07-30 16:10:32.000 BH1-1544-1-Ld		321	9	364	10	7.97669	0.23679	0.004090	0.004172	0.125794	761	22	734	135			1.24	761	22
BH1	2020-07-30 16:11:11.000 BH1-709-1-Ld		246	4	277	17	7.92100	0.24328	0.004355	0.004151	0.154546	766	22	732	136			1.86	766	22
BH1	2020-07-30 16:13:55.000 BH1-709B-1-Ld		251	5	285	15	7.91355	0.24105	0.004304	0.004151	0.150508	779	23	730	140			1.50	779	23
BH1	2020-07-30 16:15:15.000 BH1-709C-1-Ld		219	5	240	15	7.959874	0.23760	0.004565	0.004058	0.124883	763	23	811	143			5.00	763	23
BH1	2020-07-30 16:17:15.000 BH1-692-1-Ld		130	2	111	2	7.971016	0.27011	0.006382	0.004767	0.303121	762	24	801	152			4.88	762	24
BH1	2020-07-30 16:18:54.000 BH1-670B-1-Ld		116	1	91	1	7.916348	0.252689	0.005889	0.004349	0.182464	767	23	761	106			5.41	767	23
BH1	2020-07-30 16:20:00.000 BH1-648-1-Ld		47	1	39	1	7.965560	0.300282	0.006553	0.006424	0.186646	767	23	791	206			3.64	767	23
BH1	2020-07-30 16:22:04.000 BH1-648B-1-Ld		156	1	141	1	8.117774	0.31840	0.007500	0.006216	0.171791	767	23	803	195			1.74	767	23
BH1	2020-07-30 16:23:11.000 BH1-423B-1-Ld		146	3	125	3	8.11182	0.28276	0.006490	0.006216	0.171791	774	23	794	224			1.24	774	23
BH1	2020-07-30 16:32:51.000 BH1-423-1-Ld		199	5	165	5	8.20927	0.25105	0.006494	0.004837	0.207804	773	21	747	143			1.36	773	21
BH1	2020-07-30 17:02:11.000 BH1-384-1-Ld		48	2	27	2	7.514428	0.301373	0.007373	0.007073	0.212921	805	29	807	203			16.21	805	29
BH1	2020-07-30 17:02:10.000 BH1-763-1-Ld		573	7	309	9	6.351496	0.27083	0.147526	0.016786	0.028930	943	37	1217	315			93.31	943	37
BH1	2020-07-30 17:03:48.000 BH1-747-1-Ld		42	1	34	1	4.779377	0.18246	0.083440	0.008856	0.160520	1226	42	1239	160			4.14	1226	42
BH1	2020-07-30 17:05:30.000 BH1-655-1-Ld		1322	40	498	13	6.78983	0.23246	0.081847	0.005077	0.016182	803	28	1241	112			31.8	803	28
BH1	2020-07-30 17:14:15.000 BH1-643-1-Ld		402	17	226	14	5.804486	0.19461	0.07485	0.006434	0.118118	1025	32	1081	123			5.17	1025	32
BH1	2020-07-30 17:18:02.000 BH1-601-1-Ld		25	1	20	1	6.89985	0.22018	0.07485	0.007489	0.128531	1086	41	1043	204			5.34	1086	41
BH1	2020-07-30 17:17:46.000 BH1-505-1-Ld		129	2	45	2	6.125	0.15120	0.08731	0.00718	0.128531	1251	37	1371	126			8.78	1251	37
BH1	2020-07-30 17:19:25.000 BH1-446-1-Ld		974	24	2144	69	7.577399	0.21605	0.081400	0.004749	0.331341	799	21	1222	115			34.58	799	21
BH1	2020-07-30 17:21:07.000 BH1-392-1-Ld		49	3	37	3	6.82329	0.20523	0.086351	0.005456	0.20552	1232	32	1176	205			10.46	1232	32
BH1	2020-07-30 17:22:51.000 BH1-178-1-Ld		179	10	45	1	5.429986	0.154281	0.075815	0.004409	0.210696	1090	29	1089	117			0.03	1090	29
BH1	2020-07-30 17:24:29.000 BH1-172-1-Ld		255	4	118	2	2.942177	0.09068	0.13124	0.00470	0.210696	1886	50	2073	103			9.01	2073	103
BH1	2020-07-30 17:26:12.000 BH1-152-1-Ld		98	4	98	5	8.125246	0.28150	0.060408	0.004843	0.310931	748	25	809	156			7.45	748	25
BH1	2020-07-30 17:35:55.000 BH1-111-1-Ld		252	7	2	2	4.82382	0.18250	0.079482	0.004632	0.20552	1262	45	1410	127			10.46	1262	45
BH1	03/12/2018 (2) 13:50:02.000 BH1-1		272	6	90	5	4.70808	0.17390	0.080638	0.006780	0.274660	1242	40	1242	139			0.03	1242	40
BH1	03/12/2018 (2) 13:52:08.000 BH1-2		120	16	118	5	4.402286	0.25121	0.080200	0.004004	0.088564	1319	65	1201	110			9.81	1319	65
BH1	03/12/2018 (2) 13:53:09.000 BH1-3		426	14	438	4	5.418143	0.21806	0.080638	0.006780	0.274660	1242	40	1242	139			0.03	1242	40
BH1	03/12/2018 (2) 13:56:14.000 BH1-4		206	10	101	10	4.159734	0.154139	0.091900	0.006638	0.216050	1389	45	1465	113			5.17	1389	45
BH1	03/12/2018 (2) 13:58:15.000 BH1-5		198	11	83	5	3.289474	0.107990	0.104800	0.005396	0.194390	1711	48	1710	93			0.06	1710	93
BH1	03/12/2018 (2) 14:00:18.000 BH1-6		230	4	230	4	4.332131	0.159738	0.084780	0.004218	0.159738	1212	34	1212	99			0.26	1212	99
BH1	03/12/2018 (2) 14:02:20.000 BH1-7		76	2	34	1	5.153720	0.244176	0.101400	0.009028	0.254290	1074	42	1649	157			14.86	1074	42
BH1	03/12/2018 (2) 14:04:23.000 BH1-8		261	4	261	4	3.032483	0.057130	0.080638	0.006780	0.274660	1242	40	1242	139			0.03	1242	40
BH1	03/12/2018 (2) 14:06:26.000 BH1-9		122	5	12	1	5.455517	0.210233	0.074900	0.007138	0.303940	1085	39	1085	183			1.88	1085	39
BH1	03/12/2018 (2) 14:08:31.000 BH1-10		217	8	76	3	5.833803	0.18250	0.091900	0.004902	0.171791	1053	31	1057	125			3.97	1053	31
BH1	03/12/2018 (2) 14:10:33.000 BH1-11		106	2	106	2	5.218149	0.229149	0.080638	0.006780	0.274660	1242	40	1242	139			0.03	1242	40
BH1	03/12/2018 (2) 14:13:36.000 BH1-12		92	4	71	4	4.083713	0.105400	0.077700	0.007344	0.171400	1168	42	1127	180			1.35	1168	42
BH1	03/12/2018 (2) 14:14:36.000 BH1-13		68	1	31	2	4.302618	0.159738	0.080638	0.006780	0.274660	1242	40	1242	139			0.26	1242	40
BH1	03/12/2018 (2) 14:16:48.000 BH1-14		255	9	91	4	3.995586	0.159999	0.105500	0.006510	0.093335	1664	66	1722	95			3.38	1722	95
BH1	03/12/2018 (2) 14:18:50.000 BH1-15		822	37	266	11	5.814823	0.19044	0.074900	0.003098	0.183800	1057	29	1005	82			0.78	1057	29
BH1	03/12/2018 (2) 14:20:54.000 BH1-16		212	3	139	3	4.780081	0.139412	0.080638	0.006780	0.274660	1242	40	1242	139			0.26	1242	40
BH1	03/12/2018 (2) 14:22:54.000 BH1-17		151	8	73	5	6.006066	0.238751	0.075100	0.006302	0.190850	985	35	1070	161			7.99	985	35
BH1	03/12/2018 (2) 14:24:59.000 BH1-18		221	9	114	4	5.644481	0.199885	0.091900	0.005226	0.151620	1083	35	1102	132			1.17	1083	35
BH1	03/12/2018 (2) 14:26:59.000 BH1-19		99	4	99	4	3.838023	0.113000	0.080638	0.006780	0.274660	1242	40	1242	139			0.26	1242	40
BH1	03/12/2018 (2) 14:29:06.000 BH1-20		220	10	121	5	4.955401	0.17232	0.078300	0.004966	0.279600	1185	37	1154	122			2.27	1154	37
BH1	03/12/2018 (2) 14:31:06.000 BH1-21		444	2	444	2	4.807862	0.159738	0.080638	0.006780	0.274660	1242	40	1242	139			0.26	1242	40
BH1	03/12/2018 (2) 14:34:14.000 BH1-22		155	7	71	3	4.191115	0.141788	0.080638	0.006510	0.091855	1379	41	1349	122			2.29	1379	41
BH1	03/12/2018 (2) 14:36:49.000 BH1-23		84	12	92	31	7.092199	0.279735	0.082000	0.010704	0.335540	800	81	1319	227			35.51	800	81
BH1	03/12/2018 (2) 14:37:52.000 BH1-24		93	4	93	4	4.577440	0.164194	0.080638	0.006780	0.274660	1242	40	1242	139			0.26	1242	40
BH1	03/12/2018 (2) 14:39:57.000 BH1-25		189	18	83	10	4.073320	0.139538	0.088400	0.005028	0.100310	1415	42	1391	117			1.415	1415	42
BH1	03/12/2018 (2) 14:41:58.000 BH1-26		364	5	215	4	4.517605	0.137350	0.084000	0.003892	0.026997	1275	34	1267	237			2.57	1275	34
BH1	03/12/2018 (2) 14:43:58.000 BH1																			

# Appendix D.2 continued.

Sample ID	Analysis date-time	Spot ID <sup>1</sup>	218U conc.			232Th conc.			Isotopic Ratios			Assessment Ages - Calculated in Excel <sup>2</sup> (k)			Dose <sup>3</sup> (7.6-6.7)/7.6 "100"	Preferred age (50% Ma conf.)	
			ppm	±1σ	217Th/218U	ppm	±1σ	217Th/218U	238U/235U	232Th/235U	238U/235U	232Th/235U	238U/235U	232Th/235U			
BH1	03/12/2018 (2) 20:44:23.00	BH1-149	129	9	1.1	3.190911	0.104620	0.105660	0.050812	0.339622	1755	49	1174	99	1.77	1.74	
BH1	03/12/2018 (2) 20:46:30.00	BH1-150	129	9	87	5.515720	0.256346	0.079000	0.050880	0.110850	1074	44	1171	142	8.21	1074	
BH1	03/12/2018 (2) 20:48:30.00	BH1-151	240	22	101	2.884338	0.094292	0.111100	0.050620	0.407940	1919	53	1181	86	2.07	1881	
BH1	03/12/2018 (2) 20:50:37.00	BH1-152	39	1	29	1.403399	0.049100	0.083500	0.051700	0.119350	519	26	1195	246	9.52	1195	
BH1	03/12/2018 (2) 20:52:38.00	BH1-153	618	17	438	7.427897	0.182985	0.091000	0.030720	0.170270	1250	43	1446	77	13.33	1250	
BH1	03/12/2018 (2) 20:54:44.00	BH1-154	278	14	13	3.103810	0.110510	0.103100	0.050810	0.110070	1200	42	1485	162	3.88	1485	
BH1	03/12/2018 (2) 20:56:44.00	BH1-155	643	15	13	5.000510	0.165110	0.075500	0.040410	0.005345	1175	33	1081	104	8.66	1175	
BH1	03/12/2018 (2) 20:58:50.00	BH1-156	77	3	97	3.282320	0.245183	0.078000	0.030830	0.140520	1118	46	1146	200	2.48	1118	
BH1	03/12/2018 (2) 21:00:56.00	BH1-157	68	2	12	1.292040	0.102814	0.103100	0.050810	0.110070	980	26	1195	80	2.64	1195	
BH1	03/12/2018 (2) 21:03:01.00	BH1-158	738	30	266	18	6.093845	0.175759	0.076960	0.030432	0.328630	980	26	1110	88	27.50	980
BH1	03/12/2018 (2) 21:05:01.00	BH1-159	598	52	589	61	5.536186	0.282125	0.087300	0.040546	0.234930	1104	51	1266	98	10.21	1104
BH1	03/12/2018 (2) 21:07:01.00	BH1-160	190	78	384	1	1.246530	0.085500	0.050610	0.127430	569	20	1206	236	4.64	1206	
BH1	03/12/2018 (2) 21:09:08.00	BH1-161	296	18	153	9	5.924171	0.244828	0.081700	0.040734	0.274310	1006	37	1237	155	18.71	1006
BH1	03/12/2018 (2) 21:11:13.00	BH1-162	172	9	71	9	5.173296	0.191127	0.084600	0.050690	0.121000	1129	38	1383	236	12.52	1129
BH1	03/12/2018 (2) 21:13:13.00	BH1-163	92	11	81	14	10.309280	1.268998	0.115000	0.016300	0.427920	597	63	1879	236	68.24	597
BH1	03/12/2018 (2) 21:15:48.00	BH1-164	63	3	39	2	5.687978	0.347670	0.083300	0.030936	0.216820	1056	36	1217	139	20.00	1056
BH1	03/12/2018 (2) 21:17:57.00	BH1-165	172	8	2	5.88821	0.214000	0.050810	0.050810	0.204400	1016	43	1041	158	4.40	1016	
BH1	03/12/2018 (2) 21:19:57.00	BH1-166	217	10	85	4	4.399472	0.151733	0.090400	0.050810	0.170770	1320	41	1403	103	7.87	1320
BH1	03/12/2018 (2) 21:22:08.00	BH1-167	164	6	83	4	4.496483	0.180680	0.084000	0.030936	0.272090	1295	41	1297	123	4.22	1295
BH1	03/12/2018 (2) 21:24:10.00	BH1-168	1107	7	403	3	5.687135	0.167135	0.075300	0.060696	0.094789	1066	39	1076	104	9.04	1066
BH1	03/12/2018 (2) 21:26:47.00	BH1-169	156	6	158	6	4.746084	0.293937	0.084000	0.050788	0.159410	1233	66	1442	120	12.71	1233
BH1	03/12/2018 (2) 21:28:43.00	BH1-170	262	13	148	18	5.390836	0.189858	0.075400	0.050428	0.126550	1097	34	1105	108	6.70	1097
BH1	03/12/2018 (2) 21:30:54.00	BH1-171	309	47	213	45	9.515008	0.647139	0.103000	0.080806	0.309270	642	39	1629	156	60.20	642
BH1	03/12/2018 (2) 21:32:58.00	BH1-172	269	10	234	11	3.107220	0.111399	0.118300	0.050676	0.338850	1799	55	1860	88	3.32	1860
BH1	03/12/2018 (2) 21:35:00.00	BH1-173	72	5	72	1	4.819329	0.292827	0.077000	0.060824	0.030710	992	43	1050	209	6.44	992
BH1	03/12/2018 (2) 21:37:05.00	BH1-174	433	29	204	15	5.191288	0.176629	0.078000	0.040460	0.123350	1135	34	1146	101	10.97	1135
BH1	03/12/2018 (2) 21:39:07.00	BH1-175	7	447	4	4.598856	0.173745	0.075300	0.030434	0.071783	1268	43	1276	103	6.27	1268	
BH1	03/12/2018 (2) 22:01:11.00	BH1-176	1176	56	345	17	5.291236	0.142498	0.079000	0.030380	0.154450	1128	28	1171	176	3.70	1128
BH1	03/12/2018 (2) 22:03:18.00	BH1-177	253	19	144	11	4.892368	0.170787	0.080700	0.050514	0.210840	1199	37	1213	118	11.30	1199
BH1	03/12/2018 (2) 22:05:26.00	BH1-178	12	90	12	4	4.903100	0.130854	0.084000	0.050810	0.110070	1187	41	1487	110	2.84	1187
BH1	03/12/2018 (2) 22:07:26.00	BH1-179	251	10	7	0	1.828154	0.053608	0.031400	0.060698	0.181000	2813	55	2813	55	14.1	2813
BH1	03/12/2018 (2) 22:09:32.00	BH1-180	147	10	69	6	4.988406	0.292378	0.090000	0.050758	0.307903	1187	61	1444	141	17.83	1187
BH1	03/12/2018 (2) 22:11:31.00	BH1-181	86	7	4	5.88821	0.244628	0.071000	0.030936	0.217100	1194	40	1246	104	10.64	1194	
BH1	03/12/2018 (2) 22:13:39.00	BH1-182	145	25	502	18	5.767013	0.251700	0.121000	0.031940	0.441230	1031	40	1085	260	48.07	1031
BH1	03/12/2018 (2) 22:15:39.00	BH1-183	21	7	88	7	7.123296	0.230833	0.083300	0.050810	0.110070	1187	41	1487	110	2.84	1187
BH1	03/12/2018 (2) 22:17:47.00	BH1-184	155	10	90	7	3.727211	0.148823	0.101100	0.050512	0.081119	1719	65	1794	85	3.62	1719
BH1	03/12/2018 (2) 22:19:44.00	BH1-185	73	2	36	2	4.208822	0.191569	0.097000	0.030460	0.094177	1353	54	1566	156	13.58	1353
BH1	03/12/2018 (2) 22:21:51.00	BH1-186	19	149	9	4	4.028700	0.109810	0.071000	0.050810	0.110070	1187	41	1487	110	2.84	1187
BH1	03/12/2018 (2) 22:23:51.00	BH1-187	426	13	208	21	4.339084	0.133100	0.087400	0.033748	0.250000	1365	38	1569	81	1.31	1365
BH1	03/12/2018 (2) 22:25:51.00	BH1-188	23	15	2	4.767028	0.243786	0.085000	0.050810	0.110070	1187	41	1487	110	2.84	1187	
BH1	03/12/2018 (2) 22:28:00.00	BH1-189	494	25	359	20	4.187605	0.215272	0.147600	0.050752	0.065882	1381	61	2533	70	45.50	1381
BH1	03/12/2018 (2) 22:30:00.00	BH1-190	108	5	129	8	5.517298	0.230815	0.090600	0.031032	0.466980	1143	45	1447	204	20.49	1143
BH1	03/12/2018 (2) 22:32:03.00	BH1-191	122	10	122	0	5.916041	0.171723	0.074000	0.050998	0.171723	1170	35	1170	35	10.66	1170
BH1	03/12/2018 (2) 22:34:48.00	BH1-192	170	14	244	20	4.488275	0.169227	0.081500	0.060610	0.320450	1302	43	1317	134	11.5	1302
BH1	03/12/2018 (2) 22:36:53.00	BH1-193	610	10	409	13	7.942812	0.605945	0.091200	0.050746	0.021247	764	55	1473	95	48.10	764
BH1	03/12/2018 (2) 22:38:54.00	BH1-194	24	97	17	4	4.866976	0.236689	0.081500	0.041936	0.297800	1187	41	1487	110	2.84	1187
BH1	03/12/2018 (2) 22:40:58.00	BH1-195	64	3	41	1	3.061849	0.124986	0.112000	0.030840	0.102320	1822	63	1831	138	0.50	1831
BH1	03/12/2018 (2) 22:42:59.00	BH1-196	71	11	71	1	3.418827	0.148499	0.071000	0.050810	0.110070	1187	41	1487	110	2.84	1187
BH1	03/12/2018 (2) 22:45:05.00	BH1-197	138	8	39	3	3.118181	0.117626	0.119000	0.030618	0.043171	1888	63	1813	101	14.99	1813
BH1	03/12/2018 (2) 22:47:05.00	BH1-198	123	10	45	3	4.230118	0.181229	0.089000	0.060880	0.087547	1368	51	1403	142	2.40	1368
BH1	03/12/2018 (2) 22:49:11.00	BH1-199	40	2	20	2	5.400020	0.162594	0.074000	0.050810	0.110070	1187	41	1487	110	2.84	1187
BH1	03/12/2018 (2) 23:01:11.00	BH1-200	162	20	116	17	7.071236	0.551567	0.107400	0.060948	0.339880	853	58	1755	161	51.42	853
BH1	13/07/2020 (2) 12:49:08.00	BH1-1390	854	12	892	9	8.124797	0.261514	0.066600	0.020252	0.274600	748	22	824	78	9.23	748
BH1	13/07/2020 (2) 12:51:44.00	BH1-1391	2913	13	1	4.819329	0.292827	0.077000	0.060824	0.030710	992	43	1050	209	6.44	992	
BH1	13/07/2020 (2) 12:52:10.00	BH1-1392	385	6	495	3	8.839322	0.309343	0.073700	0.030744	0.279760	691	22	802	93	12.34	691
BH1	13/07/2020 (2) 12:53:46.00	BH1-1393	625	10	625	0	5.057695	0.162594	0.074000	0.050810	0.110070	1187	41	1487	110	2.84	1187
BH1	13/07/2020 (2) 12:55:08.00	BH1-2.5	233	7	105	5	4.537374	0.151211	0.087400	0.030382	0.379200	1280	37	1362	84	6.02	1280
BH1	13/07/2020 (2) 12:56:59.00	BH1-2.6	253	3	127	1	5.533919	0.190175	0.077600	0.030952	0.397710	1104	35	1186	99	2.80	1104
BH1	13/07/2020 (2) 12:58:03.00	BH1-2.7	182	2	85	2	4.767028	0.243786	0.085000	0.050810	0.110070	1187	41	1487	110	2.84	1187
BH1	13/07/2020 (2) 12:59:34.00	BH1-2.8	466	7	160	3	4.024445	0.130884	0.094800	0.030396	0.489130	1431	40	1523	71	10.68	1431
BH1	13/07/2020 (2) 13:02:24.00	BH1-2.9	195	17	195	0	4.658872	0.145451	0.083300	0.030470	0.117100	941	38	1159	86	18.82	941
BH1	13/07/2020 (2) 13:04:01.00	BH1-2.10	322														

# Appendix D.2 continued.

Sample ID	Analysis date-time	Spot ID <sup>1</sup>	218U conc.			232Th conc.			Isotopic Ratios			Assessment Ages - Calculated in August 8, 2014			Diseq <sup>2</sup>		Preferred age (50% Ma conf.)	
			ppm	±1σ	±2σ	ppm	±1σ	±2σ	238U/238Pu	230P/230Th	230P/232Th	230P/230Th	230P/232Th	230P/230Th	230P/232Th	230P/230Th		230P/232Th
BH1	13/07/2002 (2) 16:30:00	BH1-2127	141	3	106	3	3.548482	0.259385	0.095900	0.004390	0.311220	1602	97	2647	81	0.74	1654	81
BH1	13/07/2002 (2) 16:34:00	BH1-2128	202	4	104	2	2.152833	0.154292	0.171000	0.005020	0.133340	2459	138	1567	55	4.18	2567	55
BH1	13/07/2002 (2) 16:36:00	BH1-2129	395	8	5	5.664480	0.478000	0.000000	0.004390	0.133340	2459	138	1567	55	4.18	2567	55	
BH1	13/07/2002 (2) 16:37:00	BH1-2130	389	8	402	8	4.001601	0.134746	0.091400	0.003428	0.187800	1438	42	1454	71	1.11	1438	42
BH1	13/07/2002 (2) 16:40:57	BH1-2131	99	3	73	2	4.701477	0.179133	0.087200	0.004454	0.208260	1243	41	1261	103	1.44	1243	41
BH1	13/07/2002 (2) 16:42:00	BH1-2132	124	3	3	3.310118	0.180449	0.000000	0.003428	0.187800	1438	42	1454	71	1.11	1438	42	
BH1	13/07/2002 (2) 16:46:29	BH1-2133	126	2	80	1	5.205623	0.179988	0.077000	0.004654	0.161630	1133	35	1138	116	0.49	1133	35
BH1	13/07/2002 (2) 16:48:00	BH1-2134	491	50	340	49	3.408487	0.273190	0.000000	0.003788	0.165630	1003	38	1003	38	0.84	1003	38
BH1	13/07/2002 (2) 16:49:00	BH1-2135	351	16	152	8	5.844735	0.188167	0.078900	0.003378	0.193770	1106	33	1109	68	0.59	1109	68
BH1	13/07/2002 (2) 16:51:00	BH1-2136	301	3	412	4	2.145462	0.153181	0.140300	0.005106	0.265500	2466	138	2468	54	0.34	2468	54
BH1	13/07/2002 (2) 16:52:00	BH1-2137	759	8	171	2	5.954162	0.167077	0.003900	0.003778	0.121720	1089	31	1089	31	0.55	1089	31
BH1	13/07/2002 (2) 16:55:58	BH1-2138	84	4	102	4	3.298153	0.246008	0.108000	0.006600	0.145410	1707	103	1765	100	1.07	1765	100
BH1	13/07/2002 (2) 16:56:00	BH1-2139	83	2	57	2	3.170577	0.226252	0.109300	0.004886	0.130940	1767	103	1787	80	1.11	1787	80
BH1	13/07/2002 (2) 16:56:46	BH1-2140	185	2	3	3.488700	0.246155	0.101200	0.004514	0.129600	1646	98	1645	80	0.00	1645	80	
BH1	13/07/2002 (2) 16:58:31	BH1-2141	87	7	262	10	2.405002	0.169565	0.186700	0.005734	0.254480	2241	126	2713	51	1.37	2713	51
BH1	13/07/2002 (2) 16:59:53	BH1-2142	202	5	89	3	4.098361	0.137396	0.091000	0.003762	0.147300	1407	41	1469	75	0.46	1469	75
BH1	13/07/2002 (2) 17:01:21	BH1-2143	119	4	55	2	5.804818	0.120486	0.073000	0.006600	0.345400	1045	35	1068	119	2.11	1068	119
BH1	13/07/2002 (2) 17:02:52	BH1-2144	129	3	111	3	4.098361	0.137396	0.091000	0.004136	0.188140	1407	42	1467	84	3.76	1467	84
BH1	13/07/2002 (2) 17:04:00	BH1-2145	802	7	470	5	7.022145	0.237413	0.077000	0.002994	0.202600	809	26	1189	73	26.47	850	26
BH1	13/07/2002 (2) 17:05:49	BH1-2146	207	5	105	5	3.899377	0.183955	0.094900	0.004238	0.206700	1475	45	1521	84	3.29	1475	84
BH1	13/07/2002 (2) 17:07:30	BH1-2147	379	8	526	23	5.334034	0.205620	0.112900	0.004558	0.301710	1071	35	1846	72	4.19	1071	35
BH1	13/07/2002 (2) 17:08:00	BH1-2148	254	181	4	4.039883	0.171220	0.085400	0.003778	0.212550	1192	37	1146	82	11.46	1192	37	
BH1	13/07/2002 (2) 17:14:56	BH1-2149	176	1	68	1	5.497256	0.197597	0.077000	0.004758	0.340880	1407	35	1143	118	5.78	1077	35
BH1	13/07/2002 (2) 17:16:25	BH1-2150	478	8	113	2	3.984064	0.130474	0.091000	0.003230	0.284900	1444	41	1456	67	0.86	1444	41
BH1	13/07/2002 (2) 17:17:51	BH1-2151	122	3	38	1	4.020036	0.144417	0.091000	0.004412	0.209100	1406	45	1446	95	1.54	1446	95
BH1	13/07/2002 (2) 17:19:24	BH1-2152	273	7	148	4	5.090994	0.168279	0.081700	0.003734	0.277160	1160	34	1238	88	6.24	1238	88
BH1	13/07/2002 (2) 17:20:51	BH1-2153	49	1	67	1	4.697342	0.160606	0.096000	0.003972	0.333460	1427	49	1499	117	4.84	1427	117
BH1	13/07/2002 (2) 17:22:00	BH1-2154	345	6	7	7.026840	0.240341	0.000000	0.003428	0.187800	1438	42	1454	71	1.11	1438	42	
BH1	13/07/2002 (2) 17:28:21	BH1-2155	169	5	73	2	2.051703	0.146271	0.146600	0.005512	0.271700	2559	142	2523	59	1.44	2523	59
BH1	13/07/2002 (2) 17:29:06	BH1-2156	40	1	34	1	5.328862	0.242189	0.084000	0.007328	0.288770	1127	41	1206	166	6.52	1127	41
BH1	13/07/2002 (2) 17:31:27	BH1-2157	371	2	3	3.584146	0.159029	0.077000	0.004600	0.148900	1200	42	1200	42	0.50	1200	42	
BH1	13/07/2002 (2) 17:32:55	BH1-2158	100	2	38	1	4.118616	0.147473	0.091000	0.004736	0.342860	1401	42	1467	96	4.18	1467	96
BH1	13/07/2002 (2) 17:34:21	BH1-2159	241	7	101	2	4.021215	0.161027	0.091000	0.004736	0.342860	1401	42	1467	96	4.18	1467	96
BH1	13/07/2002 (2) 17:35:52	BH1-2160	100	2	66	1	4.875200	0.175962	0.085000	0.004618	0.247620	1203	38	1335	101	9.92	1203	38
BH1	13/07/2002 (2) 17:37:31	BH1-2161	425	8	243	3	3.915427	0.128899	0.096000	0.003472	0.103850	1466	42	1499	69	2.20	1466	42
BH1	13/07/2002 (2) 17:38:51	BH1-2162	598	2	152	4	4.039883	0.171220	0.085400	0.003778	0.212550	1192	37	1146	82	11.46	1192	37
BH1	13/07/2002 (2) 17:40:20	BH1-2163	195	5	101	3	4.147677	0.141444	0.000000	0.003996	0.282020	1392	41	1420	84	1.97	1392	41
BH1	13/07/2002 (2) 17:41:49	BH1-2164	152	3	109	2	5.211097	0.182970	0.077000	0.004066	0.212450	1132	35	1179	100	3.99	1132	35
BH1	13/07/2002 (2) 17:42:00	BH1-2165	294	4	4	4.546124	0.193940	0.000000	0.003428	0.187800	1438	42	1454	71	1.11	1438	42	
BH1	13/07/2002 (2) 17:44:47	BH1-2166	62	2	87	3	5.307856	0.204764	0.077000	0.005740	0.171220	1113	38	1129	103	0.68	1113	38
BH1	13/07/2002 (2) 17:46:10	BH1-2167	110	2	95	2	3.602623	0.254100	0.101200	0.004942	0.408140	1578	93	1662	88	5.06	1662	88
BH1	13/07/2002 (2) 17:47:45	BH1-2168	418	11	49	1	4.428212	0.191420	0.091000	0.004148	0.213820	1314	40	1369	80	3.98	1314	40
BH1	13/07/2002 (2) 17:49:10	BH1-2169	166	3	56	1	4.428212	0.191420	0.091000	0.004148	0.213820	1314	40	1369	80	3.98	1314	40
BH1	13/07/2002 (2) 17:50:40	BH1-2170	138	2	138	2	3.394809	0.153480	0.000000	0.003428	0.187800	1438	42	1454	71	1.11	1438	42
BH1	13/07/2002 (2) 17:52:10	BH1-2171	340	6	226	4	4.770982	0.156878	0.083800	0.003576	0.147610	1227	36	1287	83	4.69	1227	83
BH1	13/07/2002 (2) 17:53:39	BH1-2172	332	3	102	1	4.734848	0.157470	0.083900	0.003478	0.135480	1235	36	1289	79	4.20	1235	36
BH1	13/07/2002 (2) 17:55:13	BH1-2173	322	3	109	2	4.504895	0.191527	0.000000	0.003778	0.212550	1192	37	1146	82	11.46	1192	37
BH1	13/07/2002 (2) 17:56:38	BH1-2174	445	7	253	4	4.192872	0.183516	0.077000	0.004154	0.408140	1578	93	1662	88	5.06	1662	88
BH1	13/07/2002 (2) 17:58:07	BH1-2175	274	6	121	3	3.397003	0.237391	0.105200	0.003804	0.312650	1676	98	1737	66	2.76	1737	66
BH1	13/07/2002 (2) 17:59:38	BH1-2176	134	2	134	2	4.479426	0.131864	0.000000	0.004154	0.408140	1578	93	1662	88	5.06	1662	88
BH1	13/07/2002 (2) 18:00:39	BH1-2177	95	2	56	1	2.726281	0.197475	0.113800	0.005076	0.280220	2014	117	2014	117	0.71	2014	117
BH1	13/07/2002 (2) 18:01:13	BH1-2178	896	28	886	32	5.482456	0.167492	0.085000	0.003230	0.408200	1080	33	1315	68	1.85	1080	33
BH1	13/07/2002 (2) 18:02:40	BH1-2179	28	1	49	1	4.039883	0.171220	0.085400	0.003778	0.212550	1192	37	1146	82	11.46	1192	37
BH1	13/07/2002 (2) 18:03:09	BH1-2180	250	11	192	10	2.117298	0.149937	0.113000	0.005836	0.317200	2494	138	2618	55	4.04	2618	55
BH1																		

# Appendix D.2 continued.

Sample ID	Analysis date-time	Spot ID <sup>1</sup>	238U conc.			232Th conc.			Isotopic Ratios			Assessment Ages - Calculated in August 8, 2014			Diseq <sup>2</sup>		Preferred age, 1500 Ma conf.			
			ppm	±1σ	232Th/238U	ppm	±1σ	232Th/238U	±1σ	238U/235U	±1σ	232Th/232Pa	±1σ	238U/235U	±1σ	232Th/232Pa		±1σ	238U/235U	±1σ
3407819-D	17/03/2020 (3) 17:05:11.00	3407819-D-46	139	1	96	1	96	1	7.45913	0.29217	0.12070	0.004214	0.278220	783	28	2667	62	60.18	783	28
3407819-D	17/03/2020 (3) 16:00:06.00	3407819-D-27	96	2	139	3	139	3	5.21374	0.18525	0.10700	0.005414	0.008824	1131	37	2747	47	58.83	1131	37
3407819-D	17/03/2020 (3) 20:02:00.00	3407819-D-118	221	5	81	5	81	5	4.14075	0.13198	0.12060	0.005312	0.219660	631	23	2834	45	51.37	1273	63
3407819-D	17/03/2020 (3) 20:09:23.00	3407819-D-118	140	6	207	17	207	17	4.85053	0.16281	0.14990	0.005098	0.142240	1272	42	2344	58	45.74	1272	42
3407819-D	17/03/2020 (3) 19:15:44.00	3407819-D-95	167	2	107	2	107	2	5.13486	0.29943	0.10000	0.005400	0.521210	1112	57	2955	80	43.15	1112	57
3407819-D	17/03/2020 (3) 19:15:44.00	3407819-D-114	184	8	139	9	139	9	4.88485	0.17212	0.14000	0.005102	0.174210	1298	67	3128	97	41.76	1298	67
3407819-D	17/03/2020 (3) 17:31:31.00	3407819-D-59	156	1	55	1	55	1	4.65982	0.17137	0.13930	0.004786	0.227030	1253	42	2088	65	39.97	1253	42
3407819-D	17/03/2020 (3) 19:28:00.00	3407819-D-113	175	2	61	2	61	2	4.59377	0.23670	0.11700	0.005076	0.165280	1270	65	2600	59	38.37	1270	65
3407819-D	17/03/2020 (3) 15:11:00.00	3407819-D-94	35	2	48	4	48	4	5.79121	0.22132	0.10590	0.003992	0.091615	1026	36	1654	55	37.22	1026	36
3407819-D	17/03/2020 (3) 16:28:30.00	3407819-D-37	134	5	112	5	112	5	5.69310	0.18045	0.10250	0.003850	0.312070	1049	31	1659	70	37.15	1049	31
3407819-D	17/03/2020 (3) 15:45:15.00	3407819-D-21	170	2	84	2	84	2	4.33025	0.12100	0.12200	0.005414	0.155280	1226	50	1833	61	34.31	1226	50
3407819-D	17/03/2020 (3) 17:41:31.00	3407819-D-60	51	1	60	2	60	2	6.09385	0.23895	0.09120	0.005024	0.124110	980	35	1460	105	32.44	980	35
3407819-D	17/03/2020 (3) 16:07:08.00	3407819-D-26	214	2	309	3	309	3	5.91361	0.21619	0.09200	0.002944	0.165500	1007	34	1471	59	31.52	1007	34
3407819-D	17/03/2020 (3) 15:25:24.00	3407819-D-128	128	2	141	7	141	7	3.93930	0.12820	0.12730	0.003846	0.141480	1398	73	1988	62	26.69	1398	73
3407819-D	17/03/2020 (3) 17:04:50.00	3407819-D-45	74	2	85	4	85	4	3.90231	0.24549	0.12740	0.004048	0.510770	1470	83	2052	56	28.88	1470	83
3407819-D	17/03/2020 (3) 17:16:41.00	3407819-D-51	163	2	121	1	121	1	4.08163	0.24820	0.11800	0.004660	0.462280	1413	77	2925	71	26.63	1413	77
3407819-D	17/03/2020 (3) 19:15:39.00	3407819-D-80	181	2	203	2	203	2	5.42054	0.12438	0.10200	0.004046	0.138680	1022	40	1473	102	25.28	1022	40
3407819-D	17/03/2020 (3) 16:15:33.00	3407819-D-29	157	4	182	6	182	6	3.89865	0.19048	0.11100	0.004060	0.215170	1374	54	1374	54	25.64	1374	54
3407819-D	17/03/2020 (3) 19:28:00.00	3407819-D-113	72	3	69	3	69	3	4.09621	0.15432	0.11420	0.004134	0.212850	1408	48	1867	66	24.59	1408	48
3407819-D	17/03/2020 (3) 20:19:14.00	3407819-D-123	264	4	168	3	168	3	3.97406	0.25291	0.11700	0.004340	0.589650	1448	83	1907	67	24.20	1448	83
3407819-D	17/03/2020 (3) 17:18:43.00	3407819-D-52	85	1	105	2	105	2	4.28265	0.16430	0.10900	0.004380	0.289810	1383	47	1782	73	24.08	1383	47
3407819-D	17/03/2020 (3) 19:28:00.00	3407819-D-101	132	2	159	2	159	2	4.02765	0.17766	0.11400	0.004080	0.189450	1420	57	1863	65	23.24	1420	57
3407819-D	17/03/2020 (3) 19:27:41.00	3407819-D-102	202	6	182	3	182	3	3.86104	0.33645	0.11330	0.004806	0.866630	1485	114	2884	105	21.19	1485	114
3407819-D	17/03/2020 (3) 20:21:12.00	3407819-D-126	146	7	149	3	149	3	4.77531	0.18786	0.10510	0.003902	0.30580	1355	54	1715	68	21.01	1355	54
3407819-D	17/03/2020 (3) 20:21:12.00	3407819-D-127	175	2	201	2	201	2	4.77721	0.18786	0.10510	0.003902	0.30580	1355	54	1715	68	21.01	1355	54
3407819-D	17/03/2020 (3) 17:08:46.00	3407819-D-47	63	1	44	1	44	1	3.75938	0.28714	0.11730	0.005646	0.619190	1521	103	1915	86	20.59	1521	103
3407819-D	17/03/2020 (3) 17:08:46.00	3407819-D-48	93	17	117	17	117	17	4.18162	0.21745	0.11610	0.003856	0.191660	1387	69	1985	59	18.82	1387	69
3407819-D	17/03/2020 (3) 17:08:46.00	3407819-D-43	74	1	117	1	117	1	4.02768	0.14169	0.11020	0.004004	0.141690	1433	45	1802	66	20.48	1433	45
3407819-D	17/03/2020 (3) 19:51:34.00	3407819-D-109	263	14	130	15	130	15	2.12825	0.13337	0.24050	0.006110	0.147070	2488	129	3120	40	20.00	2488	129
3407819-D	17/03/2020 (3) 19:51:34.00	3407819-D-110	86	4	64	1	64	1	4.24786	0.18436	0.11300	0.004046	0.160640	1386	44	1946	66	19.86	1386	44
3407819-D	17/03/2020 (3) 17:38:29.00	3407819-D-62	87	3	115	5	115	5	4.18813	0.14388	0.10480	0.004096	0.214130	1377	43	1710	81	19.45	1377	43
3407819-D	17/03/2020 (3) 15:31:19.00	3407819-D-17	112	5	132	12	132	12	3.43834	0.23405	0.11930	0.003980	0.177060	1366	89	1941	90	18.29	1366	89
3407819-D	17/03/2020 (3) 19:19:47.00	3407819-D-98	135	4	88	3	88	3	5.42054	0.20047	0.08100	0.003112	0.177010	1092	38	1400	70	19.52	1092	38
3407819-D	17/03/2020 (3) 19:19:47.00	3407819-D-102	234	28	268	32	268	32	3.69660	0.22327	0.11330	0.004046	0.138680	1022	40	1473	102	25.28	1022	40
3407819-D	17/03/2020 (3) 19:53:35.00	3407819-D-110	125	1	153	1	153	1	3.77645	0.21843	0.11180	0.004236	0.065011	1514	78	1828	69	17.86	1828	69
3407819-D	17/03/2020 (3) 17:36:36.00	3407819-D-56	113	4	298	6	298	6	3.79506	0.23429	0.11900	0.003918	0.293960	1508	83	1813	59	18.46	1508	83
3407819-D	17/03/2020 (3) 19:53:35.00	3407819-D-110	125	1	153	1	153	1	3.77645	0.21843	0.11180	0.004236	0.065011	1514	78	1828	69	17.86	1828	69
3407819-D	17/03/2020 (3) 20:21:15.00	3407819-D-124	298	11	384	9	384	9	3.77385	0.28071	0.11060	0.003812	0.141280	1515	103	1809	63	15.21	1809	63
3407819-D	17/03/2020 (3) 18:26:05.00	3407819-D-81	157	18	86	17	86	17	3.22950	0.14207	0.10700	0.004458	0.326140	1496	50	1763	76	16.19	1496	50
3407819-D	17/03/2020 (3) 18:26:05.00	3407819-D-89	220	5	500	8	500	8	3.94950	0.23290	0.08540	0.004080	0.406770	1529	47	1324	98	12.24	1529	47
3407819-D	17/03/2020 (3) 20:07:22.00	3407819-D-117	140	4	92	3	92	3	1.81974	0.10910	0.26000	0.007020	0.223990	2824	137	3282	42	13.86	3282	42
3407819-D	17/03/2020 (3) 17:42:29.00	3407819-D-65	78	3	100	3	100	3	3.61660	0.22375	0.10420	0.004046	0.216370	1465	78	1699	83	17.79	1699	83
3407819-D	17/03/2020 (3) 17:56:29.31	3407819-D-61	355	35	254	27	254	27	3.65828	0.20555	0.10270	0.004354	0.014187	1462	67	1673	78	13.32	1462	67
3407819-D	17/03/2020 (3) 17:58:25.00	3407819-D-67	125	3	161	3	161	3	3.65893	0.23838	0.10700	0.004094	0.204950	1538	88	1794	68	13.16	1794	68
3407819-D	17/03/2020 (3) 18:26:05.00	3407819-D-81	157	18	86	17	86	17	3.22950	0.14207	0.10700	0.004458	0.326140	1496	50	1763	76	16.19	1496	50
3407819-D	17/03/2020 (3) 19:21:48.00	3407819-D-99	68	1	101	1	101	1	3.489184	0.20702	0.11360	0.004172	0.310180	1465	84	1857	66	12.32	1857	66
3407819-D	17/03/2020 (3) 18:04:18.00	3407819-D-70	226	5	83	2	83	2	3.10103	0.20775	0.11230	0.004046	0.14320	1608	84	1836	65	12.46	1836	65
3407819-D	17/03/2020 (3) 18:04:18.00	3407819-D-111	111	3	127	3	127	3	3.65449	0.19176	0.11230	0.004172	0.21910	1390	62	1900	66	12.24	1900	

# Appendix D.2 continued.

Sample ID	Analysis date-time	Spot ID <sup>1</sup>	Z18U conc.			Z17Th conc.			Isotopic Ratios			Account Ages - Calculated in August R. Ma			Dose <sup>2</sup>		Preferred age 1500 <sup>3</sup>				
			ppm	±1σ	ppm	±1σ	ppm	±1σ	Z18U/Z206Pb	Z17Th/Z206Pb	±1σ PRCP	±1σ	Z18U/Z206Pb	±1σ PRCP	Z17Th/Z206Pb	±1σ PRCP	(78-6/92) %	Ma	±1σ		
OD1719	2020-07-30 17:46:53.000	OD1719-14	1	1	7.80224	0.23550	0.00001	0.31194	764	65	0.00014	0.01194	764	65	0.00014	0.01194	764	65	0.00014	0.01194	
OD1719	2020-07-30 16:58:40.000	OD1719-13	1	1	6.25401	0.32150	0.00672	0.00613	0.26306	737	27	0.011	0.26306	737	27	0.011	0.26306	737	27	0.011	
OD1719	2020-07-30 16:58:40.000	OD1719-12	1	1	4.7	0.78930	0.13471	0.00102	0.00816	0.21713	768	31	0.011	0.21713	768	31	0.011	0.21713	768	31	0.011
OD1719	10/05/2019 (01:07:25.69)	OD1719-20	49	3	4.29941	0.19866	0.00400	0.00832	0.17800	1104	37	0.011	0.17800	1104	37	0.011	0.17800	1104	37	0.011	
OD1719	10/05/2019 (01:04:22.00)	OD1719-30	317	9	4.54545	0.38015	0.15100	0.03200	0.05120	1282	90	0.011	0.05120	1282	90	0.011	0.05120	1282	90	0.011	
OD1719	10/05/2019 (01:04:22.00)	OD1719-29	285	4	4.33333	0.25555	0.00000	0.03200	0.21600	1095	66	0.011	0.21600	1095	66	0.011	0.21600	1095	66	0.011	
OD1719	10/05/2019 (01:04:23.00)	OD1719-104	65	4	4.30662	0.19920	0.11100	0.01002	0.43230	1346	54	0.011	0.43230	1346	54	0.011	0.43230	1346	54	0.011	
OD1719	10/05/2019 (01:04:27.00)	OD1719-22	53	2	3.66975	0.17978	0.15340	0.00998	0.57470	1554	65	0.011	0.57470	1554	65	0.011	0.57470	1554	65	0.011	
OD1719	10/05/2019 (01:05:05.00)	OD1719-64	141	2	7.13302	0.27078	0.00000	0.01200	0.21600	1150	59	0.011	0.21600	1150	59	0.011	0.21600	1150	59	0.011	
OD1719	10/05/2019 (01:05:17.50)	OD1719-51	499	53	5.0	2.70124	0.25744	0.09200	0.01180	0.21830	1042	42	0.011	0.21830	1042	42	0.011	0.21830	1042	42	0.011
OD1719	10/05/2019 (01:05:38.00)	OD1719-59	311	18	108	4.73701	0.18890	0.04800	0.00896	0.14170	1235	45	0.011	0.14170	1235	45	0.011	0.14170	1235	45	0.011
OD1719	10/05/2019 (01:05:43.00)	OD1719-45	94	8	5.85486	0.21772	0.00400	0.00832	0.21600	1104	37	0.011	0.21600	1104	37	0.011	0.21600	1104	37	0.011	
OD1719	10/05/2019 (01:05:58.70)	OD1719-46	216	9	5.20028	0.30925	0.09300	0.01580	0.31900	1134	59	0.011	0.31900	1134	59	0.011	0.31900	1134	59	0.011	
OD1719	10/05/2019 (01:06:07.00)	OD1719-25	120	5	5.13028	0.20388	0.00400	0.01200	0.21600	1150	59	0.011	0.21600	1150	59	0.011	0.21600	1150	59	0.011	
OD1719	10/05/2019 (01:06:12.10)	OD1719-48	236	9	219	8.12754	0.15530	0.10740	0.01048	0.20240	1374	44	0.011	0.20240	1374	44	0.011	0.20240	1374	44	0.011
OD1719	10/05/2019 (01:06:48.00)	OD1719-64	89	4	4.40917	0.26509	0.10300	0.01500	0.31020	1318	68	0.011	0.31020	1318	68	0.011	0.31020	1318	68	0.011	
OD1719	10/05/2019 (01:07:47.10)	OD1719-113	137	6	3.78921	0.15992	0.11800	0.00676	0.20390	1529	56	0.011	0.20390	1529	56	0.011	0.20390	1529	56	0.011	
OD1719	10/05/2019 (01:08:14.00)	OD1719-65	189	11	118	6.54148	0.21081	0.08770	0.00754	0.39370	1099	38	0.011	0.39370	1099	38	0.011	0.39370	1099	38	0.011
OD1719	10/05/2019 (01:08:40.96)	OD1719-64	240	9	146	5.51269	0.22683	0.08800	0.00750	0.01697	1075	41	0.011	0.01697	1075	41	0.011	0.01697	1075	41	0.011
OD1719	10/05/2019 (01:08:42.00)	OD1719-61	208	5	152	5.52754	0.26592	0.08400	0.01130	0.24120	1072	45	0.011	0.24120	1072	45	0.011	0.24120	1072	45	0.011
OD1719	10/05/2019 (01:08:43.00)	OD1719-86	364	8	523	3.47701	0.14813	0.12250	0.00760	0.34920	1630	59	0.011	0.34920	1630	59	0.011	0.34920	1630	59	0.011
OD1719	10/05/2019 (01:08:55.00)	OD1719-103	216	8	88	5.65639	0.20388	0.00400	0.00588	0.11870	1049	34	0.011	0.11870	1049	34	0.011	0.11870	1049	34	0.011
OD1719	10/05/2019 (01:09:27.10)	OD1719-56	523	11	137	5.62664	0.30928	0.07400	0.00978	0.11570	955	42	0.011	0.11570	955	42	0.011	0.11570	955	42	0.011
OD1719	10/05/2019 (01:09:30.00)	OD1719-68	207	29	208	3.59452	0.13670	0.11900	0.01082	0.17820	1725	64	0.011	0.17820	1725	64	0.011	0.17820	1725	64	0.011
OD1719	10/05/2019 (01:09:45.00)	OD1719-14	437	13	228	5.22460	0.21873	0.08500	0.01042	0.42190	1129	42	0.011	0.42190	1129	42	0.011	0.42190	1129	42	0.011
OD1719	10/05/2019 (01:09:45.00)	OD1719-60	960	63	612	5.45537	0.12844	0.11200	0.00774	0.07940	1549	46	0.011	0.07940	1549	46	0.011	0.07940	1549	46	0.011
OD1719	10/05/2019 (01:09:52.36)	OD1719-127	212	10	73	6.52745	0.26491	0.07400	0.00628	0.16020	919	34	0.011	0.16020	919	34	0.011	0.16020	919	34	0.011
OD1719	10/05/2019 (01:09:58.00)	OD1719-134	78	2	49	5.07094	0.16782	0.08600	0.01770	0.21680	1160	68	0.011	0.21680	1160	68	0.011	0.21680	1160	68	0.011
OD1719	10/05/2019 (01:10:03.00)	OD1719-62	142	4	145	4.50768	0.10978	0.08100	0.01650	0.21600	1066	68	0.011	0.21600	1066	68	0.011	0.21600	1066	68	0.011
OD1719	10/05/2019 (01:10:25.20)	OD1719-55	994	54	134	5.43738	0.39267	0.00200	0.01840	0.07203	1088	68	0.011	0.07203	1088	68	0.011	0.07203	1088	68	0.011
OD1719	10/05/2019 (01:10:37.00)	OD1719-16	101	4	78	5.89975	0.29207	0.07100	0.01042	0.19830	1009	44	0.011	0.19830	1009	44	0.011	0.19830	1009	44	0.011
OD1719	10/05/2019 (01:10:53.00)	OD1719-22	292	2	292	5.41221	0.24944	0.00400	0.01130	0.24810	1111	20	0.011	0.24810	1111	20	0.011	0.24810	1111	20	0.011
OD1719	10/05/2019 (01:10:59.00)	OD1719-65	52	1	28	3.17596	0.12489	0.12280	0.00916	0.29640	1766	59	0.011	0.29640	1766	59	0.011	0.29640	1766	59	0.011
OD1719	10/05/2019 (01:11:03.00)	OD1719-93	93	4	93	5.12147	0.12489	0.12280	0.00916	0.29640	1766	59	0.011	0.29640	1766	59	0.011	0.29640	1766	59	0.011
OD1719	10/05/2019 (01:11:19.00)	OD1719-52	163	7	85	4.40347	0.18909	0.09300	0.00916	0.31220	1119	49	0.011	0.31220	1119	49	0.011	0.31220	1119	49	0.011
OD1719	10/05/2019 (01:11:40.59)	OD1719-49	100	4	39	2.23243	0.12945	0.14400	0.01680	0.14100	2386	110	0.011	0.14100	2386	110	0.011	0.14100	2386	110	0.011
OD1719	10/05/2019 (01:11:42.00)	OD1719-90	350	6	62	5.07584	0.20770	0.08100	0.01070	0.21600	1160	68	0.011	0.21600	1160	68	0.011	0.21600	1160	68	0.011
OD1719	10/05/2019 (01:11:50.10)	OD1719-47	155	6	54	3.34830	0.18149	0.09900	0.01192	0.22690	1456	60	0.011	0.22690	1456	60	0.011	0.22690	1456	60	0.011
OD1719	10/05/2019 (01:12:03.00)	OD1719-65	765	15	765	5.45537	0.12844	0.11200	0.00774	0.07940	1549	46	0.011	0.07940	1549	46	0.011	0.07940	1549	46	0.011
OD1719	10/05/2019 (01:12:18.36)	OD1719-127	212	10	73	6.52745	0.26491	0.07400	0.00628	0.16020	919	34	0.011	0.16020	919	34	0.011	0.16020	919	34	0.011
OD1719	10/05/2019 (01:12:36.00)	OD1719-134	78	2	49	5.07094	0.16782	0.08600	0.01770	0.21680	1160	68	0.011	0.21680	1160	68	0.011	0.21680	1160	68	0.011
OD1719	10/05/2019 (01:12:43.00)	OD1719-62	142	4	145	4.50768	0.10978	0.08100	0.01650	0.21600	1066	68	0.011	0.21600	1066	68	0.011	0.21600	1066	68	0.011
OD1719	10/05/2019 (01:12:52.20)	OD1719-55	994	54	134	5.43738	0.39267	0.00200	0.01840	0.07203	1088	68	0.011	0.07203	1088	68	0.011	0.07203	1088	68	0.011
OD1719	10/05/2019 (01:13:07.00)	OD1719-16	101	4	78	5.89975	0.29207	0.07100	0.01042	0.19830	1009	44	0.011	0.19830	1009	44	0.011	0.19830	1009	44	0.011
OD1719	10/05/2019 (01:13:13.00)	OD1719-22	292	2	292	5.41221	0.24944	0.00400	0.01130	0.24810	1111	20	0.011	0.24810	1111	20	0.011	0.24810	1111	20	0.011
OD1719	10/05/2019 (01:13:19.00)	OD1719-65	52	1	28	3.17596	0.12489	0.12280	0.00916	0.29640	1766	59	0.011	0.29640	1766	59	0.011	0.29640	1766	59	0.01





# Appendix D.2 continued.

Sample ID	Analysis date-time	Spot ID <sup>1</sup>	218U conc.		232Th conc.		Isotopic Ratios						Age-corrected (Calculated in Isotope R Ma)				Diseq <sup>2</sup>		Preferred age 1500 Ma conf.	
			ppm	±2σ	ppm	±2σ	238U/206Pb	±2σ	235U/238U	±2σ	207Pb/206Pb	±2σ	207Pb/238U	±2σ	207Pb/206Pb	±2σ	207Pb/238U	±2σ	Ma	±2σ
0407B18	14/07/2020 (h) 00:56:48.00	0407B18 - 2.104	323	7	321	7	4.273504	0.142306	0.017700	0.003314	0.311030	1355	39	1315	71	1.4	1.35	99	1.4	
0407B18	14/07/2020 (h) 01:05:55.00	0407B18 - 2.107	323	9	327	8	3.003003	0.213367	0.115000	0.004100	0.405480	1853	108	1879	64	1.37	1879	64	1.37	
0407B18	14/07/2020 (h) 01:57:57.00	0407B18 - 2.139	284	7	192	4	3.475883	0.202739	0.101600	0.003812	0.245250	1630	98	1653	66	1.39	1653	66	1.39	
0407B18	13/07/2020 (h) 20:35:35.00	0407B18 - 2.43	295	8	347	13	4.370384	0.149895	0.026900	0.003400	0.254510	1376	37	1292	75	1.25	1272	37	1.25	
0407B18	13/07/2020 (h) 23:28:33.00	0407B18 - 2.54	391	8	677	14	3.137596	0.224619	0.109300	0.004086	0.178830	1766	103	1787	67	1.19	1787	67	1.19	
0407B18	13/07/2020 (h) 23:40:37.00	0407B18 - 2.41	182	2	76	1	4.255483	0.145190	0.088000	0.004160	0.334680	1366	41	1382	89	1.40	1386	41	1.40	
0407B18	10/05/2019 (h) 02:06:06.00	0407B18 - 38	96	5	85	3	4.789272	0.182947	0.081600	0.007312	0.085653	1222	41	1235	168	1.03	1222	41	1.03	
0407B18	13/07/2020 (h) 22:39:03.00	0407B18 - 2.30	107	4	50	3	3.778862	0.275369	0.095100	0.004702	0.332190	1514	92	1529	91	1.01	1529	91	1.01	
0407B18	13/07/2020 (h) 23:03:32.00	0407B18 - 2.45	440	4	438	3	3.833886	0.262841	0.097900	0.003812	0.203482	1494	63	1509	69	0.99	1498	69	0.99	
0407B18	13/07/2020 (h) 23:16:47.00	0407B18 - 2.15	277	3	147	2	3.965107	0.131185	0.091800	0.003636	0.328780	1450	42	1462	74	0.86	1450	42	0.86	
0407B18	14/07/2020 (h) 00:28:21.00	0407B18 - 2.88	128	1	104	1	5.678502	0.203662	0.074500	0.004450	0.236600	1046	34	1054	118	0.80	1046	34	0.80	
0407B18	10/05/2019 (h) 00:46:40.00	0407B18 - 65	274	19	337	30	3.975969	0.193197	0.092300	0.005644	0.338770	1480	63	1491	184	0.77	1489	63	0.77	
0407B18	14/07/2020 (h) 00:50:49.00	0407B18 - 2.100	211	3	221	1	3.113325	0.217351	0.110600	0.004012	0.219560	1796	103	1809	65	0.71	1809	65	0.71	
0407B18	13/07/2020 (h) 23:38:40.00	0407B18 - 2.37	216	3	120	2	4.122973	0.145660	0.097900	0.003958	0.123270	1370	42	1379	83	0.68	1370	42	0.68	
0407B18	14/07/2020 (h) 02:17:35.00	0407B18 - 2.146	444	12	301	10	4.006410	0.136308	0.091000	0.003520	0.267860	1436	43	1446	73	0.64	1436	43	0.64	
0407B18	14/07/2020 (h) 02:32:23.00	0407B18 - 2.156	451	9	215	4	4.861489	0.158877	0.087000	0.003014	0.278390	1206	35	1213	73	0.61	1206	35	0.61	
0407B18	14/07/2020 (h) 02:36:16.00	0407B18 - 2.114	324	7	159	4	3.998401	0.112726	0.091000	0.003212	0.484990	1439	42	1448	75	0.61	1439	42	0.61	
0407B18	14/07/2020 (h) 03:12:47.00	0407B18 - 2.145	331	3	130	1	3.146680	0.174452	0.078200	0.003864	0.289500	1145	34	1151	34	0.56	1145	34	0.56	
0407B18	13/07/2020 (h) 23:57:27.00	0407B18 - 2.40	90	1	45	1	4.490346	0.160278	0.084300	0.003290	0.384790	1296	41	1303	118	0.54	1296	41	0.54	
0407B18	14/07/2020 (h) 02:36:50.00	0407B18 - 2.159	397	5	173	4	3.892654	0.127443	0.091100	0.003322	0.388310	1441	41	1448	73	0.47	1441	41	0.47	
0407B18	10/05/2019 (h) 00:33:05.00	0407B18 - 9	427	7	421	5	4.533734	0.216641	0.083700	0.010674	0.062781	1280	54	1285	231	0.40	1280	54	0.40	
0407B18	14/07/2020 (h) 02:34:48.00	0407B18 - 2.168	244	3	224	7	3.306615	0.212308	0.113600	0.004172	0.165670	1851	108	1857	66	0.32	1857	66	0.32	
0407B18	14/07/2020 (h) 03:01:02.00	0407B18 - 2.81	381	8	502	10	3.591954	0.098100	0.098100	0.003462	0.212740	1583	65	1588	65	0.26	1588	65	0.26	
0407B18	14/07/2020 (h) 03:05:18.00	0407B18 - 2.101	221	5	240	4	3.920021	0.139447	0.091000	0.003942	0.160660	1465	42	1469	80	0.26	1465	42	0.26	
0407B18	10/05/2019 (h) 00:39:75	0407B18 - 31	160	3	84	3	2.021886	0.074774	0.117900	0.006978	0.293190	2095	78	2095	91	0.17	2095	91	0.17	
0407B18	13/07/2020 (h) 23:50:11.00	0407B18 - 2.38	117	1	78	1	3.958828	0.183731	0.091400	0.004518	0.131740	1452	44	1464	92	0.15	1452	44	0.15	
0407B18	13/07/2020 (h) 23:50:37.00	0407B18 - 2.152	257	5	57	2	3.120773	0.105330	0.096900	0.004092	0.170580	1791	51	1792	67	0.04	1792	67	0.04	
0407B18	13/07/2020 (h) 23:50:37.00	0407B18 - 2.24	149	5	112	2	3.926188	0.135559	0.091800	0.004236	0.176980	1463	44	1462	86	0.02	1463	44	0.02	
0407B18	14/07/2020 (h) 01:31:14.00	0407B18 - 2.123	183	5	85	3	3.951427	0.128899	0.091900	0.004038	0.381950	1466	42	1465	82	-0.12	1466	42	-0.12	
0407B18	10/05/2019 (h) 02:44:40.00	0407B18 - 2.1	343	13	144	7	3.189763	0.149744	0.074600	0.003022	0.114710	1758	127	1755	127	0.17	1755	127	0.17	
0407B18	14/07/2020 (h) 01:32:46.00	0407B18 - 2.122	315	3	155	2	4.183505	0.149289	0.087000	0.003514	0.279390	1333	39	1331	39	-0.18	1333	39	-0.18	
0407B18	14/07/2020 (h) 02:11:38.00	0407B18 - 2.183	244	7	174	5	2.988240	0.200742	0.114300	0.004386	0.216090	1872	108	1888	68	-0.19	1888	68	-0.19	
0407B18	14/07/2020 (h) 02:15:44.00	0407B18 - 2.155	298	2	121	1	4.844609	0.145380	0.089800	0.003316	0.295800	1336	39	1334	74	0.21	1336	39	0.21	
0407B18	10/05/2019 (h) 00:34:56.00	0407B18 - 4	125	2	62	2	3.889517	0.144356	0.092200	0.006844	0.216540	1475	47	1471	130	0.29	1475	47	0.29	
0407B18	13/07/2020 (h) 02:33:00.00	0407B18 - 2.159	262	1	125	2	3.180662	0.120291	0.091000	0.004260	0.203100	1762	47	1757	71	0.17	1762	47	0.17	
0407B18	13/07/2020 (h) 02:37:00.00	0407B18 - 2.53	236	6	243	7	3.812383	0.131351	0.091800	0.003636	0.154280	1467	43	1462	74	0.33	1467	43	0.33	
0407B18	13/07/2020 (h) 23:31:31.00	0407B18 - 2.156	141	1	238	8	3.932245	0.122219	0.097500	0.004502	0.204990	1583	46	1576	86	0.40	1576	86	0.40	
0407B18	14/07/2020 (h) 01:03:43.00	0407B18 - 2.117	203	1	1	1	3.801903	0.167774	0.117900	0.006978	0.293190	2095	78	2095	91	0.60	2095	91	0.60	
0407B18	14/07/2020 (h) 01:03:43.00	0407B18 - 2.85	131	1	74	1	3.852080	0.269942	0.092000	0.004512	0.458720	1488	88	1479	81	-0.40	1488	88	-0.40	
0407B18	14/07/2020 (h) 01:03:43.00	0407B18 - 2.118	590	3	382	6	3.862485	0.123488	0.091400	0.003348	0.191690	1488	41	1475	65	0.65	1488	41	0.65	
0407B18	14/07/2020 (h) 01:07:22.00	0407B18 - 2.108	286	4	80	2	5.750411	0.194371	0.073500	0.003270	0.312770	1034	31	1027	88	0.65	1034	31	0.65	
0407B18	13/07/2020 (h) 23:47:51.00	0407B18 - 2.67	63	1	120	1	1.968892	0.140168	0.177200	0.007644	0.202820	2648	146	2626	71	-0.82	2626	71	-0.82	
0407B18	13/07/2020 (h) 23:48:35.00	0407B18 - 2.07	682	2	302	2	4.866812	0.142036	0.086100	0.003022	0.114710	1758	127	1755	127	0.17	1755	127	0.17	
0407B18	13/07/2020 (h) 23:48:55.00	0407B18 - 2.61	125	3	46	1	3.926188	0.135559	0.091000	0.004422	0.241780	1463	44	1463	44	-1.03	1463	44	-1.03	
0407B18	14/07/2020 (h) 01:03:50.00	0407B18 - 2.109	113	2	67	1	1.823817	0.128657	0.139200	0.006904	0.364150	2816	153	2786	54	-1.07	2786	54	-1.07	
0407B18	14/07/2020 (h) 01:03:51.00	0407B18 - 2.117	239	5	298	6	5.073507	0.168398	0.092000	0.003760	0.222050	1480	34	1486	94	0.20	1480	34	0.20	
0407B18	13/07/2020 (h) 23:21:31.00	0407B18 - 2.118	92	1	283	5	4.859086	0.127907	0.097000	0.004794	0.171760	1206	38	1189	115	-1.49	1206	38	-1.49	
0407B18	10/05/2019 (h) 01:44:40.00	0407B18 - 2.1	343	13	144	7	3.189763	0.149744	0.074600	0.003022	0.114710	1758	127	1755	127	0.17	1755	127	0.17	
0407B18	14/07/2020 (h) 01:53:33.00	0407B18 - 2.136	274	3	195	2	3.086420													







# Appendix D5.3 continued.

Sample ID	Analysis Date	Site ID	Elevation (m)	Soil Sample Data										Soil Calculations										Notes																														
				1700/1720 120/120	1710/1730 120/120	1720/1740 120/120	1730/1750 120/120	1740/1760 120/120	1750/1770 120/120	1760/1780 120/120	1770/1790 120/120	1780/1790 120/120	1790/1800 120/120	1800/1810 120/120	1810/1820 120/120	1820/1830 120/120	1830/1840 120/120	1840/1850 120/120	1850/1860 120/120	1860/1870 120/120	1870/1880 120/120	1880/1890 120/120	1890/1900 120/120																															
007078	10/05/2010	10121320	001	0.1	1.2	2.3	3.4	4.5	5.6	6.7	7.8	8.9	9.0	10.1	11.2	12.3	13.4	14.5	15.6	16.7	17.8	18.9	19.0	20.1	21.2	22.3	23.4	24.5	25.6	26.7	27.8	28.9	29.0	30.1	31.2	32.3	33.4	34.5	35.6	36.7	37.8	38.9	39.0	40.1	41.2	42.3	43.4	44.5	45.6	46.7	47.8	48.9	50.0	
007078	10/05/2010	10121330	001	0.2	1.3	2.4	3.5	4.6	5.7	6.8	7.9	8.0	9.1	10.2	11.3	12.4	13.5	14.6	15.7	16.8	17.9	18.0	19.1	20.2	21.3	22.4	23.5	24.6	25.7	26.8	27.9	28.0	29.1	30.2	31.3	32.4	33.5	34.6	35.7	36.8	37.9	38.0	39.1	40.2	41.3	42.4	43.5	44.6	45.7	46.8	47.9	48.0	49.1	50.0
007078	10/05/2010	10121340	001	0.3	1.4	2.5	3.6	4.7	5.8	6.9	7.0	8.1	9.2	10.3	11.4	12.5	13.6	14.7	15.8	16.9	17.0	18.1	19.2	20.3	21.4	22.5	23.6	24.7	25.8	26.9	27.0	28.1	29.2	30.3	31.4	32.5	33.6	34.7	35.8	36.9	37.0	38.1	39.2	40.3	41.4	42.5	43.6	44.7	45.8	46.9	47.0	48.1	49.2	50.0
007078	10/05/2010	10121350	001	0.4	1.5	2.6	3.7	4.8	5.9	6.0	7.1	8.2	9.3	10.4	11.5	12.6	13.7	14.8	15.9	16.0	17.1	18.2	19.3	20.4	21.5	22.6	23.7	24.8	25.9	26.0	27.1	28.2	29.3	30.4	31.5	32.6	33.7	34.8	35.9	36.0	37.1	38.2	39.3	40.4	41.5	42.6	43.7	44.8	45.9	46.0	47.1	48.2	49.3	50.0
007078	10/05/2010	10121360	001	0.5	1.6	2.7	3.8	4.9	5.0	6.1	7.2	8.3	9.4	10.5	11.6	12.7	13.8	14.9	15.0	16.1	17.2	18.3	19.4	20.5	21.6	22.7	23.8	24.9	25.0	26.1	27.2	28.3	29.4	30.5	31.6	32.7	33.8	34.9	35.0	36.1	37.2	38.3	39.4	40.5	41.6	42.7	43.8	44.9	45.0	46.1	47.2	48.3	49.4	50.0
007078	10/05/2010	10121370	001	0.6	1.7	2.8	3.9	4.0	5.1	6.2	7.3	8.4	9.5	10.6	11.7	12.8	13.9	14.0	15.1	16.2	17.3	18.4	19.5	20.6	21.7	22.8	23.9	24.0	25.1	26.2	27.3	28.4	29.5	30.6	31.7	32.8	33.9	34.0	35.1	36.2	37.3	38.4	39.5	40.6	41.7	42.8	43.9	44.0	45.1	46.2	47.3	48.4	49.5	50.0
007078	10/05/2010	10121380	001	0.7	1.8	2.9	3.0	4.1	5.2	6.3	7.4	8.5	9.6	10.7	11.8	12.9	13.0	14.1	15.2	16.3	17.4	18.5	19.6	20.7	21.8	22.9	23.0	24.1	25.2	26.3	27.4	28.5	29.6	30.7	31.8	32.9	33.0	34.1	35.2	36.3	37.4	38.5	39.6	40.7	41.8	42.9	43.0	44.1	45.2	46.3	47.4	48.5	49.6	50.0
007078	10/05/2010	10121390	001	0.8	1.9	3.0	3.1	4.2	5.3	6.4	7.5	8.6	9.7	10.8	11.9	12.0	13.1	14.2	15.3	16.4	17.5	18.6	19.7	20.8	21.9	22.0	23.1	24.2	25.3	26.4	27.5	28.6	29.7	30.8	31.9	32.0	33.1	34.2	35.3	36.4	37.5	38.6	39.7	40.8	41.9	42.0	43.1	44.2	45.3	46.4	47.5	48.6	49.7	50.0
007078	10/05/2010	10121400	001	0.9	2.0	3.1	3.2	4.3	5.4	6.5	7.6	8.7	9.8	10.9	11.0	12.1	13.2	14.3	15.4	16.5	17.6	18.7	19.8	20.9	21.0	22.1	23.2	24.3	25.4	26.5	27.6	28.7	29.8	30.9	31.0	32.1	33.2	34.3	35.4	36.5	37.6	38.7	39.8	40.9	41.0	42.1	43.2	44.3	45.4	46.5	47.6	48.7	49.8	50.0
007078	10/05/2010	10121410	001	1.0	2.1	3.2	3.3	4.4	5.5	6.6	7.7	8.8	9.9	10.0	11.1	12.2	13.3	14.4	15.5	16.6	17.7	18.8	19.9	20.0	21.1	22.2	23.3	24.4	25.5	26.6	27.7	28.8	29.9	30.0	31.1	32.2	33.3	34.4	35.5	36.6	37.7	38.8	39.9	40.0	41.1	42.2	43.3	44.4	45.5	46.6	47.7	48.8	49.9	50.0
007078	10/05/2010	10121420	001	1.1	2.2	3.3	3.4	4.5	5.6	6.7	7.8	8.9	9.0	10.1	11.2	12.3	13.4	14.5	15.6	16.7	17.8	18.9	19.0	20.1	21.2	22.3	23.4	24.5	25.6	26.7	27.8	28.9	29.0	30.1	31.2	32.3	33.4	34.5	35.6	36.7	37.8	38.9	39.0	40.1	41.2	42.3	43.4	44.5	45.6	46.7	47.8	48.9	49.0	50.0
007078	10/05/2010	10121430	001	1.2	2.3	3.4	3.5	4.6	5.7	6.8	7.9	8.0	9.1	10.2	11.3	12.4	13.5	14.6	15.7	16.8	17.9	18.0	19.1	20.2	21.3	22.4	23.5	24.6	25.7	26.8	27.9	28.0	29.1	30.2	31.3	32.4	33.5	34.6	35.7	36.8	37.9	38.0	39.1	40.2	41.3	42.4	43.5	44.6	45.7	46.8	47.9	48.0	49.1	50.0
007078	10/05/2010	10121440	001	1.3	2.4	3.5	3.6	4.7	5.8	6.9	7.0	8.1	9.2	10.3	11.4	12.5	13.6	14.7	15.8	16.9	17.0	18.1	19.2	20.3	21.4	22.5	23.6	24.7	25.8	26.9	27.0	28.1	29.2	30.3	31.4	32.5	33.6	34.7	35.8	36.9	37.0	38.1	39.2	40.3	41.4	42.5	43.6	44.7	45.8	46.9	47.0	48.1	49.2	50.0
007078	10/05/2010	10121450	001	1.4	2.5	3.6	3.7	4.8	5.9	6.0	7.1	8.2	9.3	10.4	11.5	12.6	13.7	14.8	15.9	16.0	17.1	18.2	19.3	20.4	21.5	22.6	23.7	24.8	25.9	26.0	27.1	28.2	29.3	30.4	31.5	32.6	33.7	34.8	35.9	36.0	37.1	38.2	39.3	40.4	41.5	42.6	43.7	44.8	45.9	46.0	47.1	48.2	49.3	50.0
007078	10/05/2010	10121460	001	1.5	2.6	3.7	3.8	4.9	5.0	6.1	7.2	8.3	9.4	10.5	11.6	12.7	13.8	14.9	15.0	16.1	17.2	18.3	19.4	20.5	21.6	22.7	23.8	24.9	25.0	26.1	27.2	28.3	29.4	30.5	31.6	32.7	33.8	34.9	35.0	36.1	37.2	38.3	39.4	40.5	41.6	42.7	43.8	44.9	45.0	46.1	47.2	48.3	49.4	50.0
007078	10/05/2010	10121470	001	1.6	2.7	3.8	3.9	4.0	5.1	6.2	7.3	8.4	9.5	10.6	11.7	12.8	13.9	14.0	15.1	16.2	17.3	18.4	19.5	20.6	21.7	22.8	23.9	24.0	25.1	26.2	27.3	28.4	29.5	30.6	31.7	32.8	33.9	34.0	35.1	36.2	37.3	38.4	39.5	40.6	41.7	42.8	43.9	44.0	45.1	46.2	47.3	48.4	49.5	50.0
007078	10/05/2010	10121480	001	1.7	2.8	3.9	4.0	4.1	5.2	6.3	7.4	8.5	9.6	10.7	11.8	12.9	13.0	14.1	15.2	16.3	17.4	18.5	19.6	20.7	21.8	22.9	23.0	24.1	25.2	26.3	27.4	28.5	29.6	30.7	31.8	32.9	33.0	34.1	35.2	36.3	37.4	38.5	39.6	40.7	41.8	42.9	43.0	44.1	45.2	46.3	47.4	48.5	49.6	50.0
007078	10/05/2010	10121490	001	1.8	2.9	3.0	4.1	4.2	5.3	6.4	7.5	8.6	9.7	10.8	11.9	12.0	13.1	14.2	15.3	16.4	17.5	18.6	19.7	20.8	21.9	22.0	23.1	24.2	25.3	26.4	27.5	28.6	29.7	30.8	31.9	32.0	33.1	34.2	35.3	36.4	37.5	38.6	39.7	40.8	41.9	42.0	43.1	44.2	45.3	46.4	47.5	48.6	49.7	50.0
007078	10/05/2010	10121500	001	1.9	3.0	3.1	4.2	4.3	5.4	6.5	7.6	8.7	9.8	10.9	11.0	12.1	13.2	14.3	15.4	16.5	17.6	18.7	19.8	20.9	21.0	22.1	23.2	24.3	25.4	26.5	27.6	28.7	29.8	30.9	31.0	32.1	33.2	34.3	35.4	36.5	37.6	38.7	39.8	40.9	41.0	42.1	43.2	44.3	45.4	46.5	47.6	48.7	49.8	50.0
007078	10/05/2010	10121510	001	2.0	3.1	3.2	4.3	4.4	5.5	6.6	7.7	8.8	9.9	10.0	11.1	12.2	13.3	14.4	15.5	16.6	17.7	18.8	19.9	20.0	21.1	22.2	23.3	24.4	25.5	26.6	27.7	28.8	29.9	30.0	31.1	32.2	33.3	34.4	35.5	36.6	37.7	38.8	39.9	40.0	41.1	42.2	43.3	44.4	45.5	46.6	47.7	48.8	49.9	50.0
007078	10/05/2010	10121520	001	2.1	3.2	3.3	4.4	4.5	5.6	6.7	7.8	8.9	9.0	10.1	11.2	12.3	13.4	14.5	15.6	16.7	17.8	18.9	19.0	20.1	21.2	22.3	23.4	24.5	25.6	26.7	27.8	28.9	29.0	30.1	31.2	32.3	33.4	34.5	35.6	36.7	37.8	38.9	39.0	40.1	41.2	42.3	43.4	44.5	45.6	46.7	47.8	48.9	49.0	50.0
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## Appendix D5.4 continued.

Metric	Method	Reference	BH1	03DTB19	04DTB19	05DTB19
			Youngest analysis	single analysis	Ludwig and Mundil, 2002	715±22
Youngest single analysis with uncertainty	single analysis	Dickinson and Gehrels, 2009	715±22	764±22	739±21	768±31
Youngest single grain (multiple analysis)	Weighted mean	Spencer et al., 2016	735±11	N/A	759±18	N/A
Youngest analysis cluster 2 or more (1σ overlap)	Weighted mean	Dickinson and Gehrels, 2009	718±18	770±16	757±10	N/A
Youngest cluster 3 or more (1σ overlap)	Weighted mean	Dickinson and Gehrels, 2009	727±11	770±16	757±10	N/A
Youngest three grains	Weighted mean	Zhang et al., 2016	723±14	770±16	749±14	N/A
Youngest cluster 2+ (2σ overlap)	Weighted mean	Coutts et al., 2019	736±17	770±16	760±16	N/A
Concordia age of youngest cluster	Concordia Age	Dehler et al., 2010	758±8 (1 rej)	770±16	760±16	N/A
Youngest Unmix population (6+ grains needed)	Isoplot unmix function	Dehler et al., 2017	729±13	N/A	746±30	N/A
Youngest graphical peak (PDP, 2 grain minimum)	Probability density plot	Dickinson and Gehrels, 2009	759	769	759	N/A
Maximum Likelihood Age	IsoplotR radial plot minimum finite mixture	Vermeesch, 2020	726±7	771±7	760±5	N/A

### Appendix D5.5: U/Pb-Lu/Hf Standards

Recommended  $^{176}\text{Hf}/^{177}\text{Hf}$  value for zircon reference materials and weighted mean standard corrected ratio obtained in each analytical session (Mud Tank primary reference material).

Zircon standard	Recommended $^{176}\text{Hf}/^{177}\text{Hf}$	Weighted mean for session 1 (3/12/2018)	Weighted mean for session 2 (09/05/2019)	Weighted mean for session 3 (17/03/2020)
<b>Mud Tank</b>	0.282505 ± 0.000044 (Woodhead and Hergt, 2005)	0.282507 ± 0.000006 (MSWD=0.47, n = 17)	0.282507 ± 0.000004 (MSWD = 0.89, n = 20)	0.282507 ± 0.000009 (MSWD=0.11, n = 20)
<b>GJ-1</b>	0.282000 ± 0.000005, (Morelet et al., 2008)	0.282027 ± 0.000008 (MSWD = 0.21, n = 16)	0.282011 ± 0.000005 (MSWD = 1.00, n = 19)	0.282013 ± 0.000005 (MSWD=0.30, n = 19)
<b>Plešovice</b>	0.282482 ± 0.000013 (Sláma et al., 2008)	0.282485 ± 0.000009 (MSWD=0.52, n = 10)	0.282473 ± 0.000006 (MSWD = 0.77, n = 8)	0.282477 ± 0.000006 (MSWD=0.55, n = 9)
<b>91500</b>	0.282306 ± 0.000311 (Woodhead et al., 2004)	0.282300 ± 0.000009 (MSWD=0.18, n = 15)	0.282303 ± 0.000006 (MSWD = 1.03, n = 18)	0.282306 ± 0.000005 (MSWD=0.59, n = 19)

Recommended ages for zircon reference materials and weighted mean standard corrected ages obtained in each analytical session.

Zircon standard	Recommended age (Ma)	1st session (03/12/2018)	2nd session (9/05/2019)	3rd session (7/03/2020)
<b>GJ-1</b>	608.5 ± 1.5 Ma (Jackson et al., 2004) 601.5 ± 0.40 Ma (Horstwood et al., 2016)	603.6 ± 0.88 Ma (MSWD=0.83, n = 20)	608.3 ± 1.4 Ma (MSWD=5.0, n = 17)	602.2 ± 0.84 Ma (MSWD=1.38, n = 19)
<b>Plešovice</b>	337.13 ± 0.37 Ma (Sláma et al., 2008)	338.4 ± 0.7 Ma (MSWD=2.0, n = 9)	-	339.5 ± 1.79 Ma (MSWD=4.5, n = 7)
<b>91500</b>	1062.4 ± 0.4 Ma (Wiedenbeck et al., 1995)	1062.0 ± 2.6 Ma (MSWD=0.54, n = 20)	1061.4 ± 3.9 Ma (MSWD=0.01, n = 20)	1062.3 ± 1.6 Ma (MSWD=0.89, n = 20)
<b>OGC</b>	3467 ± 3 Ma (Stern et al., 2009)	3465.2 ± 2.2 Ma (MSWD=0.54, n = 18)	-	3465.1 ± 4.0 Ma (MSWD=1.69, n = 14)

## Appendix D6.1: Compiled U/Pb Data Sources

U-Pb Detrital Zircon Data presented in Fig. 5

Sample Number	Area	Unit	Source
09JK08	Pocatello	Camelback Mountain Fm	Yonkee et al., 2014
20AY11	Pocatello	Geertsen Canyon Fm	Yonkee et al., 2014
66CD10	Pocatello	Geertsen Canyon Fm	Yonkee et al., 2014
05Z08	Pocatello	Camelback Mountain Fm	Yonkee et al., 2014
03Z08	Pocatello	Mutual Fm	Yonkee et al., 2014
04Z08	Pocatello	Mutual Fm	Yonkee et al., 2014
01JK08	Pocatello	Caddy Canyon Fm	Yonkee et al., 2014
65CD10	Pocatello	Browns Hole Fm	Yonkee et al., 2014
59CD10	Pocatello	Mutual Fm	Yonkee et al., 2014
57CD10	Pocatello	Caddy Canyon Fm	Yonkee et al., 2014
Sample E*	Grand Canyon	Sixtymile Fm	Karlstrom et al., 2018
Sample D*	Grand Canyon	Sixtymile Fm	Karlstrom et al., 2018
Sample C*	Grand Canyon	Sixtymile Fm	Karlstrom et al., 2018
Sample B*	Grand Canyon	Sixtymile Fm	Karlstrom et al., 2018
Sample A*	Grand Canyon	Sixtymile Fm	Karlstrom et al., 2018
Tapeats 1*	Grand Canyon	Tapeats Sandstone	Gehrels et al., 2011
Sample F*	Grand Canyon	Tapeats Sandstone	Karlstrom et al., 2018
Sample G*	Grand Canyon	Tapeats Sandstone	Karlstrom et al., 2018
Sample H*	Grand Canyon	Tapeats Sandstone	Karlstrom et al., 2018
06SA10*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
04SASC2*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
04SASC6*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
07SD3c*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
07SD3a*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
09MS1*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
08FM8b*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
08FM9*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
09SL1*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
09SL2*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
10BM216*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
10BM215*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
11LHM03*	New Mexico	Bliss Sandstone	Amato and Mack, 2012
CS12-5*	New Mexico	Van Horn Fm	Spencer et al., 2014
CS12-3*	New Mexico	Van Horn Fm	Spencer et al., 2014
MD09	C. Appalachians	Weverton Fm	Satoski, 2013
MD10	C. Appalachians	Weverton Fm	Satoski, 2013
MD11	C. Appalachians	Weverton Fm	Satoski, 2013
VA12	C. Appalachians	Weverton Fm	Satoski, 2013
VA13	C. Appalachians	Weverton Fm	Satoski, 2013
VA14	C. Appalachians	Weverton Fm	Satoski, 2013
VA15	C. Appalachians	Weverton Fm	Satoski, 2013
VA16	C. Appalachians	Weverton Fm	Satoski, 2013
Rome 1	C. Appalachians	Rome Sandstone	Thomas, 2004
Rome 2	C. Appalachians	Rome Sandstone	Thomas, 2004

\*Presented max depositional age constraints (shown in Fig. 2) are from Karlstrom et al., 2018 and shown in Karlstrom et al., 2018 supplementary tables

## Appendix D6.2: Compiled Lu/Hf Data Sources

Lu-Hf Detrital Zircon Data from late Tonian-Cambrian strata presented in Fig. 3 and 4

Sample Number	Group	Unit	Source
08LW6	Tonian-Cryogenian Strata	Little Willow Fm	Spencer et al., 2012
BCF-HF	Tonian-Cryogenian Strata	Middle Big Cottonwood Fm	Spencer et al., 2012
LW17-3	Tonian-Cryogenian Strata	Little Willow Fm	Spencer et al., 2012
Caddy	Tonian-Cryogenian Strata	Caddy Canyon Quartzite	Gehrels and Pecha, 2014
J39-A-9	Ediacaran Strata	Johnnie Fm	Howard et al., 2015
NR30	Ediacaran Strata	Stirling Quartzite	Howard et al., 2015
Mutual	Ediacaran Strata	Mutual Fm	Gehrels and Pecha, 2014
WQ	Ediacaran Strata	Wildhorse Meadows	Wooden et al., 2013
Geertsen	Cambrian Strata	Geertsen Canyon Fm	Gehrels and Pecha, 2014
Wood Canyon	Cambrian Strata	Wood Canyon Fm	Gehrels and Pecha, 2014
85NV_CL_3B	Cambrian Strata	Wood Canyon Fm	Howard et al., 2015
83WMAM_4	Cambrian Strata	Campito Fm	Howard et al., 2015
Zabriskie2	Cambrian Strata	Zabriskie Quartzite	Gehrels and Pecha, 2014
Provedora	Cambrian Strata	Provedora	Gehrels and Pecha, 2014
ZAB	Cambrian Strata	Zabriskie Quartzite	Wooden et al., 2013
ZBR	Cambrian Strata	Zabriskie Quartzite	Wooden et al., 2013
WC	Cambrian Strata	Wood Canyon Fm	Wooden et al., 2013
12COPP-03	Pikes Peak	Pikes Peak Syenogranite	Howard et al., 2015
12COPP-02	Pikes Peak	Pikes Peak Monzogranite	Howard et al., 2015
12COPP-05	Pikes Peak	Pikes Peak Quartz syenite	Howard et al., 2015
Yqs-10	Pikes Peak	Lake George quartz syenite	Guitreau et al., 2016
Ypm-9	Pikes Peak	Lake George monzogranite	Guitreau et al., 2016
Ypp-20	Pikes Peak	Pikes Peak granite	Guitreau et al., 2016
11LBM_01	Llano/Wichita	Little Hatchet Mountain Granite	Howard et al., 2015
12TXFM01	Llano/Wichita	Franklin Mountains, Stage 2	Howard et al., 2015
12TXFM02	Llano/Wichita	Franklin Mountains, Stage 4	Howard et al., 2015
12TXFM03	Llano/Wichita	Franklin Mountains, Stage 1	Howard et al., 2015
5/04 LL-1	Llano/Wichita	Escuadra Granite	Howard et al., 2015
GRLL-5	Llano/Wichita	Lone Grove pluton	Howard et al., 2015
GRLL-14	Llano/Wichita	Grape Creek pluton	Howard et al., 2015
CS12-4	Llano/Wichita	compiled in Martin et al., 2020	Mulder et al., 2018
CS12-5	Llano/Wichita	compiled in Martin et al., 2020	Mulder et al., 2019
CS12-1	Llano/Wichita	compiled in Martin et al., 2020	Mulder et al., 2020
CS12-3	Llano/Wichita	compiled in Martin et al., 2020	Mulder et al., 2021
OK-4-WL	Llano/Wichita	compiled in Martin et al., 2020	Thomas et al., 2016
OK-1-VN	Llano/Wichita	compiled in Martin et al., 2020	Thomas et al., 2016
BR-1	Grenville	Blowing Rock Gneiss	Howard et al., 2015
CG	Grenville	Corbin Gneiss	Howard et al., 2015
TX	Grenville	Toxaway Gneiss	Howard et al., 2015
WL	Grenville	Wiley Gneiss	Howard et al., 2015
WD	Grenville	Woodland Gneiss	Howard et al., 2015
13-AM-007	Grenville	compiled in Martin et al., 2020	Augland et al., 2016
13-FS-1029	Grenville	compiled in Martin et al., 2020	Augland et al., 2016
13-TC-5016	Grenville	compiled in Martin et al., 2020	Augland et al., 2016
n3422	Grenville	compiled in Martin et al., 2020	Spencer et al., 2015
n3423	Grenville	compiled in Martin et al., 2020	Spencer et al., 2015
n3427	Grenville	compiled in Martin et al., 2020	Spencer et al., 2015
CS11-1	Grenville	compiled in Martin et al., 2020	Spencer et al., 2015
CS11-13	Grenville	compiled in Martin et al., 2020	Spencer et al., 2015
CS11-18	Grenville	compiled in Martin et al., 2020	Spencer et al., 2015
CS11-20	Grenville	compiled in Martin et al., 2020	Spencer et al., 2015
CS11-3	Grenville	compiled in Martin et al., 2020	Spencer et al., 2015
CS11-6	Grenville	compiled in Martin et al., 2020	Spencer et al., 2015
CS11-9	Grenville	compiled in Martin et al., 2020	Spencer et al., 2015
KY-18-CB	Grenville	compiled in Martin et al., 2020	Thomas et al., 2017
KY-19-PR7	Grenville	compiled in Martin et al., 2020	Thomas et al., 2017
KY-21-SG	Grenville	compiled in Martin et al., 2020	Thomas et al., 2017
OH-1-SS	Grenville	compiled in Martin et al., 2020	Thomas et al., 2017
VA-1-GN	Grenville	compiled in Martin et al., 2020	Thomas et al., 2017
WV-1-PR	Grenville	compiled in Martin et al., 2020	Thomas et al., 2017

Abbreviations: FM - Formation



## Appendix D6.3: Statistical Comparison Results

Statistical results generated from DZStats-2D (Sundell and Saylor, 2021)

Sundell, K.E., and Saylor, J.E., 2021, Two-dimensional Quantitative Comparison of Density Distributions in Detrital Geochronology and Geochemistry: Geochemistry Geophysics Geosystems, doi:10.1029/2020GC009559.

DZStats2D Results

Cross-correlation

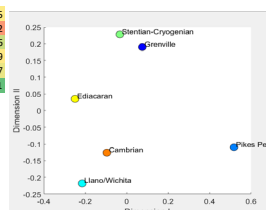
	Grenville	Pikes Peak	Llano/Wichita	Stentian-Cryogenian	Ediacaran	Cambrian
Grenville	1	0.426854753	0.274316864	0.770183729	0.474454264	0.462331331
Pikes Peak	0.426854753	1	0.062011811	0.187423138	0.078132217	0.19763872
Llano/Wichita	0.274316864	0.062011811	1	0.380173424	0.653479709	0.800640595
Stentian-Cryogenian	0.770183729	0.187423138	0.380173424	1	0.56196941	0.472054653
Ediacaran	0.474454264	0.078132217	0.653479709	0.56196941	1	0.692551412
Cambrian	0.462331331	0.19763872	0.800640595	0.472054653	0.692551412	1

MDS PLOTS

See Fig. 4

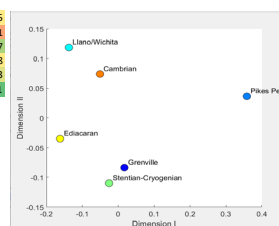
Likeness

	Grenville	Pikes Peak	Llano/Wichita	Stentian-Cryogenian	Ediacaran	Cambrian
Grenville	1	0.424673852	0.468235074	0.742896371	0.568893514	0.581816995
Pikes Peak	0.424673852	1	0.244411589	0.336260072	0.219678172	0.346164282
Llano/Wichita	0.468235074	0.244411589	1	0.500314997	0.52121186	0.723636186
Stentian-Cryogenian	0.742896371	0.336260072	0.500314997	1	0.610730274	0.567177089
Ediacaran	0.568893514	0.219678172	0.62121186	0.610730274	1	0.637237657
Cambrian	0.581816995	0.346164282	0.723636186	0.567177089	0.637237657	1



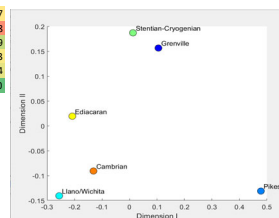
Similarity

	Grenville	Pikes Peak	Llano/Wichita	Stentian-Cryogenian	Ediacaran	Cambrian
Grenville	1	0.681305315	0.772454111	0.940079556	0.833254383	0.835563425
Pikes Peak	0.681305315	1	0.518008562	0.615254083	0.475171631	0.638635801
Llano/Wichita	0.772454111	0.518008562	1	0.790880589	0.862038612	0.927946867
Stentian-Cryogenian	0.940079556	0.615254083	0.790880589	1	0.837474096	0.831758658
Ediacaran	0.833254383	0.475171631	0.862038612	0.837474096	1	0.869925073
Cambrian	0.835563425	0.638635801	0.927946867	0.831758658	0.869925073	1



K-S Test D

	Grenville	Pikes Peak	Llano/Wichita	Stentian-Cryogenian	Ediacaran	Cambrian
Grenville	0	0.476731512	0.475021841	0.080312451	0.374636295	0.368292847
Pikes Peak	0.476731512	0	0.734338073	0.551847976	0.666269051	0.62988338
Llano/Wichita	0.475021841	0.734338073	0	0.429978063	0.298053989	0.216186869
Stentian-Cryogenian	0.080312451	0.551847976	0.429978063	0	0.313646632	0.340530273
Ediacaran	0.374636295	0.666269051	0.298053989	0.313646632	0	0.29685264
Cambrian	0.368292847	0.62988338	0.216186869	0.340530273	0.29685264	0



Kuiper Test V

	Grenville	Pikes Peak	Llano/Wichita	Stentian-Cryogenian	Ediacaran	Cambrian
Grenville	0	0.535770229	0.476568755	0.101301506	0.394890118	0.369835468
Pikes Peak	0.535770229	0	0.7394555	0.63232497	0.72219529	0.638849709
Llano/Wichita	0.476568755	0.7394555	0	0.438857807	0.34116225	0.219808536
Stentian-Cryogenian	0.101301506	0.63232497	0.438857807	0	0.327609597	0.340530273
Ediacaran	0.394890118	0.72219529	0.34116225	0.327609597	0	0.346740626
Cambrian	0.369835468	0.638849709	0.219808536	0.340530273	0.346740626	0

