Instability of the southern Canadian Shield during the late Proterozoic

Kalin T. McDannell[a,b]*, Peter K. Zeitler[a], and David A. Schneider[c]

[a]*Department of Earth and Environmental Sciences, Lehigh University, 1 W. Packer Ave. Bethlehem PA, 18015 USA*

[b]*Geological Survey of Canada, Natural Resources Canada, 3303 – 33 St NW Calgary AB, T2L 2A7 Canada*

[c]*Department of Earth & Environmental Sciences, University of Ottawa, 25 Templeton Ave., Ottawa ON, K1N 6N5 Canada*

*corresponding author: kmcdannell@gmail.com; kalin.mcdannell@canada.ca

ABSTRACT

Cratons are generally considered to comprise lithosphere that has remained tectonically quiescent for billions of years. Direct evidence for stability is mainly founded in the Phanerozoic sedimentary record and low-temperature thermochronology, but for extensive parts of Canada, earlier stability has been inferred due to the lack of an extensive rock record in both time and space. We used $^{40}Ar/^{39}Ar$ multi-diffusion domain (MDD) analysis of K-feldspar to constrain cratonic thermal histories across an intermediate (~150-350°C) temperature range in an attempt to link published high-temperature geochronology that resolves the timing of orogenesis and metamorphism with lower-temperature data suited for upper-crustal burial and unroofing histories. This work is focused on understanding the transition from Archean-Paleoproterozoic crustal growth to later intervals of stability, and how uninterrupted that record is throughout Earth's Proterozoic "Middle Age." Intermediate-temperature thermal histories of cratonic rocks at well-constrained localities within the southern Canadian Shield of North America challenge the stability worldview because our data indicate that these rocks were at elevated temperatures in the Proterozoic. Feldspars from granitic rocks collected at the surface cooled at rates of <0.5°C/Ma subsequent to orogenesis, seemingly characteristic of cratonic lithosphere, but modeled thermal histories suggest that at ca. 1.1-1.0 Ga these rocks were still near ~200°C – signaling either reheating, or prolonged residence at mid-crustal depths assuming a normal cratonic geothermal gradient. After 1.0 Ga, the regions we sampled then underwent further cooling such that they were at or near the surface (<< 60°C) in the early Paleozoic. Explaining mid-crustal residence at 1.0 Ga is challenging. A widespread, prolonged reheating history via burial is not supported by stratigraphic information, however assuming a purely monotonic cooling history requires at the very least 5 km of exhumation beginning at ca. 1.0 Ga. A possible

explanation may be found in evidence of magmatic underplating that thickened the crust, driving uplift and erosion. The timing of this underplating coincides with Mid-Continent extension, Grenville orogenesis, and assembly of the supercontinent Rodinia. $^{40}Ar/^{39}Ar$ MDD data demonstrate that this technique can be successfully applied to older rocks and fill in a large observational gap. These data also raise questions about the evolution of cratons during the Proterozoic and the nature of cratonic stability across deep time.

Keywords: craton; $^{40}Ar/^{39}Ar$; Superior Province; thermochronology; stability; underplating

1. Introduction

1.1 Cratons and stabilization

Cratons are made of ancient, tectonically quiescent continental lithosphere that has characteristically thick, chemically depleted mantle and low heat flow (e.g. Jaupart et al., 2014; Jordan, 1978; Rudnick and Nyblade, 1999). Cratons are intrinsically interesting because they record the secular evolution of the continents and a cyclic response to mantle dynamics. Moreover, cratonization and weathering of large, stable landmasses are considered integral to geo-biological evolution linked to fundamental environmental changes such as atmospheric and oceanic oxygenation and nitrification (e.g. Young, 2013 and references therein).

Here we define “stability” as resistance to internal deformation, limited magmatism, and maintenance of continental freeboard (Pollack, 1986). The established view is that most cratons underwent a period of long-term stability for most of the Proterozoic following early consolidation and exhumation at the end of the Archean. The lack of a record of significant tectonic events in the long interval between initial Archean cratonization and epeirogenic activity in the Phanerozoic is the basis for this assessment. This large time gap corresponds to the Great Unconformity, which in some places represents over 1-2 billion years of time (e.g. Karlstrom and Timmons, 2012). For the majority of the interval, time-temperature data are lacking, and there are few preserved extensive or temporally continuous sedimentary rocks in many cratonic terranes from this period, thus leaving the erosional record poorly constrained. Geochronologic study of cratons has usually been focused either on understanding cratonization and metamorphism during orogenic suturing using high-closure temperature (>500°C) methods (e.g.

Schneider et al., 2007) or alternatively, examination of more recent shallow exhumation and burial histories using methods with low closure temperatures (<100°C) (e.g. Ault et al. 2009). From a thermochronological standpoint, the 1 to 2 billion-year gap in the Proterozoic record could simply reflect the lack of studies using methods that address the 500°C to 100°C temperature window.

This period of apparent cratonic lithosphere stability between 1.7 to 0.75 Ga contributes to the idea that it represents the "Boring Billion" or "Earth's Middle Age" (Cawood and Hawkesworth, 2014 and references therein), a time of geological, biologic, and environmental stability that is attributed in part to secular cooling of the mantle and the development of a strong continental lithosphere. The Columbia supercontinent cycle from ca. 2.1-1.3 Ga (Rogers and Santosh, 2002) began this period of pervasive stability and endured until protracted breakup of Rodinia between ca. 0.86-0.57 Ga (Li et al., 2008). The 1.1 Ga orogenic belts that occur on several modern continents (Hoffman, 1988; Li et al. 2008) are the basis for primary reconstructions of the late Mesoproterozoic supercontinent Rodinia. Rodinia formed through global contractional orogenesis from ca. 1.3-1.0 Ga. Reconstructed paleogeography demonstrates that the entire southeastern margin of Laurentia experienced several stages of accretion between ca. 1.75 and 1.0 Ga, whereas the Laurentian core underwent continental-scale northeast-southwest extension, causing rifting and numerous dike swarms (e.g. Whitmeyer and Karlstrom, 2007).

1.2 Stability: record or lack thereof?

The Phanerozoic sedimentary record within continental interiors primarily establishes the direct geologic evidence for more recent stability. Cratons in particular seem to have experienced no more than ~1-3 km of vertical motion over many hundreds of millions of years during Paleozoic-Mesozoic time, based on a modest amount of stratigraphic data and low-temperature thermochronology (Ault et al., 2009; Flowers et al., 2012; Flowers et al., 2006a). Suggested mechanisms for the observed epeirogenic motions in North America include far-field lithospheric responses to subduction (Mitrovica et al., 1989) and dynamic topography induced by mantle flow (Forte et al., 2010). In central Canada, the case for stability is based mostly on what is missing: a paucity of Proterozoic orogenic activity and preserved sediments, and a lack of disturbed (i.e. thermally reset) geochronological systems.

Extensive work in the western Superior Province suggests that orogenic activity over the Kenoran interval (Kenorland; ca. 2.75 to 2.65 Ga) ceased by 2.6 Ga and transitioned shortly thereafter to stable cratonic lithosphere by 2.5 Ga, with some immediate post-orogenic sedimentation (e.g. the Huronian Supergroup; Percival et al., 2012). Internal deformation occurred during mantle plume activity in the form of the Matatchewan large igneous province (LIP) and dike swarm at ca. 2.48-2.45 Ga during proto-continental breakup (Ernst and Bleeker, 2010) and 2.2 Ga emplacement of the Nipissing Sills fed by the Ungava plume. The Kapuskasing Uplift in the western Superior Province caused local deformation that exposed the middle-lower crust at ca. 1.9 Ga (Percival and West, 1994). The 1.86 Ga Penokean Orogeny resulted in the Himalayan-scale orogen that existed along the southern Superior margin, synchronous with the Trans-Hudson Orogeny (THO; ca. 1.9-1.8 Ga) to the north and northwest (Bickford et al., 2005) leading to Laurentian assembly. Yavapai orogenesis occured to the south at ca. 1.75 Ga, leading to deposition of the Athabasca and Baraboo sequences, and other mature sandstones across Laurentia (Davidson, 2008 for review). Later, Rodinia assembly and Grenvillian orogenesis caused continental-scale extension, while concurrent hotspot magmatism produced the 1.27 Ga Mackenzie dike swarm, 1.24 Ga Sudbury dike swarm, and the 1.1 Ga Mid-Continental Rift (MCR) volcanism and Abitibi dike swarm (Ernst and Bleeker, 2010). Following these events, subtle epeirogenic motions of up to ±1-2 km characterized the Phanerozoic (Feinstein et al., 2009; Flowers et al., 2012).

The Trans-Hudson Orogen underwent an orogenic and stabilization sequence analogous to that of the earlier Kenoran Orogeny, where ca. 1.9 to 1.8 Ga granulite-facies metamorphic rocks (Alexandre et al., 2009; Flowers et al., 2008; Schneider et al., 2007; Williams and Hanmer, 2006) underlie Athabasca sediments deposited between ca. 1.7 to 1.6 Ga (Rainbird et al., 2007). This was seemingly followed by quiescence until modest Phanerozoic epeirogenesis (Ault et al., 2009; Flowers et al., 2012). Almost a billion years of thermal record is unaccounted for after initial stabilization. It is clear that certain locations on the North American craton such as Athabasca show that orogenic mid-crustal rocks were exhumed into the shallow crust soon after orogenesis (e.g. Williams and Hanmer, 2006). However, many other parts of the Canadian Shield are lacking conclusive information on post-orogenic activity in the late Archean through Proterozoic.

This paper addresses the apparent post-orogenic stability of the southern Canadian Shield during the Proterozoic using K-feldspar $^{40}Ar/^{39}Ar$ thermochronology and examines potential mechanisms to explain the thermal histories suggested by these data, including cratonic disruption near the end of the Proterozoic caused by crustal thickening and uplift due to magmatic underplating. Our conclusions suggest that parts of the Canadian Shield experienced significant exhumation beginning at ca. 1.0 Ga.

2. K-feldspar multi-diffusion domain $^{40}Ar/^{39}Ar$ thermochronology

2.1 Linking high and low-temperature thermochronometers

Potassium feldspar is a useful mineral for $^{40}Ar/^{39}Ar$ dating because of its high K content, ubiquity in felsic rocks, and stability during *in vacuo* heating (McDougall and Harrison, 1999). K-feldspar is moderately retentive of radiogenic ^{40}Ar, and low-temperature ordered K-feldspars usually exhibit complex microstructures that lead to what has been termed multi-diffusion-domain (MDD) behavior, which means that most grains of K-feldspar contain a distribution of diffusion domains that can record a range of temperatures (Lovera et al., 1989; Lovera et al., 1991). $^{40}Ar/^{39}Ar$ K-feldspar MDD analysis is able to determine continuous temperature-time (T-t) paths over the range ~150˚C to 350˚C and the shape of age spectra can distinguish between slow and rapid cooling. Continuous T-t paths can be established over this interval because the ^{39}Ar released during step-heating can be used to independently determine the sample's specific diffusion kinetics and domain-size distribution. These data can then be used to invert the observed $^{40}Ar/^{39}Ar$ age spectrum for the only remaining unknown, the thermal history. Modeling for thermal history is best done using an inverse approach to eliminate observer bias from forward modeling, and also speed up the recovery of a robust suite of thermal histories. K-feldspar $^{40}Ar/^{39}Ar$ thermochronology fills the middle-temperature range between higher temperature methods recording crystallization, such as U-Pb geochronology, and the low-temperature chronometers (apatite fission-track and (U-Th)/He) that record near-surface processes.

Within a Superior Province reference frame, our craton-interior samples are spatially bracketed by well-known histories in the northwestern craton (Athabasca) and the southeastern craton margin (Grenville; fig. 1). Samples of granitic rocks from surface exposures and near-surface drill cores were chosen for their K-feldspar content and because most of the sample sites have existing high- and low-temperature thermochronologic constraints. The supplementary material contains comprehensive information on rock samples and geochronologic context.

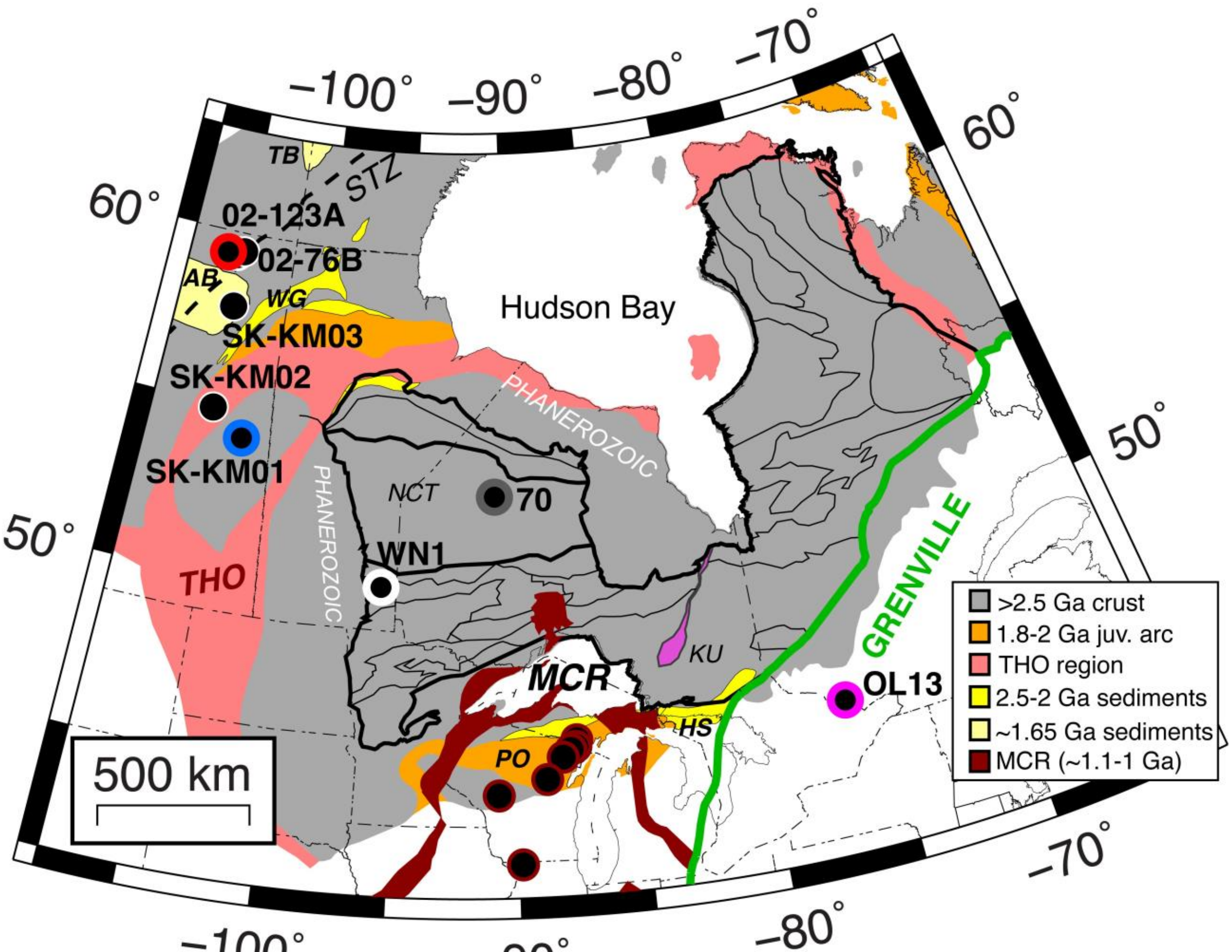

Figure 1. Map of the southern Canadian Shield modified from Whitmeyer and Karlstrom (2007) with the Superior Province (outlined by thick black line) and associated sub-terrane boundaries after Percival et al. (2006); actual western Superior craton boundary extends to the THO beneath Phanerozoic sedimentary cover. Red, blue, and violet circles correspond to samples and age spectra/thermal histories in figure 3, whereas other black dots show locations of additional K-feldspar samples we have dated by $^{40}Ar/^{39}Ar$ MDD step-heating (i.e. samples and data south of the MCR are shown in the supplement). Gray areas are >2.5 Ga Archean crust and orange regions are 2.0-1.8 Ga juvenile arcs of the Penokean orogen in the south, La Ronge to the northwest, and Torngat-Narsajuaq to the northeast. Pink shading reflects areas affected by the Trans-Hudson orogeny (THO). Darker yellow areas are 2.5-2.0 Ga post-orogenic passive

margin sediments, HS = Huronian Supergroup; WG = Wollaston Group; OG = Opswagan Group. Light yellow areas are ca. 1.6-1.7 Ga sediments of the Athabasca Basin (AB) and the Thelon Basin (TB). NCT = North Caribou Terrane; STZ = Snowbird Tectonic Zone; THO = Trans Hudson Orogen; PO = Penokean Orogen; MCR = Mid-Continent Rift (dark red shading), KU = Kapuskasing Uplift (shaded violet). Green line is Grenville Province boundary.

2.2 ^{40}Ar/^{39}Ar MDD thermochronology results

Granitic rocks from various parts of the Canadian Shield including the Western Superior Province (Ontario) and terranes affected by the THO (Saskatchewan) yield discordant ^{40}Ar/^{39}Ar age spectra that systematically droop in a "staircase" manner. In a good number of cases (fig. 2) these age spectra are interpretable as reflecting diffusional response to slow cooling or reheating, unafflicted by such phenomena as breakage of large domains or late recrystallization. Overall our results align with the ~70% success rate reported by Lovera et al. (2002) for a large database of samples. Information about ^{40}Ar/^{39}Ar dating methodology and modeling can be found in the supplementary material.

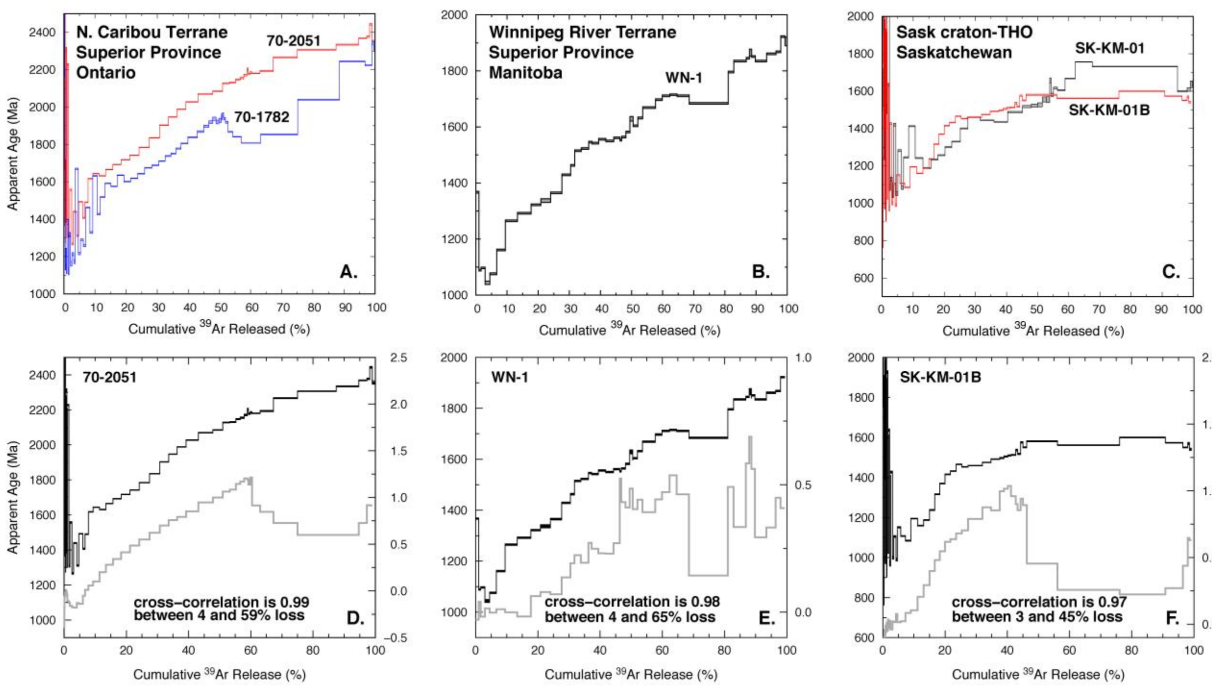

Figure 2. (A-C) Subset of step-heated ^{40}Ar/^{39}Ar age spectra for K-feldspars from the Canadian Shield. Superior Province samples #70 and WN-1 illustrate slow cooling over ca. 800-900 m.y. The Sask craton sample SK-KM-01 replicates indicate slow cooling synchronous with Superior Province samples from ca. 1.6 to ca. 1.0 Ga. Most samples have excess Ar in early release steps and experienced melting near 1050-1100°C during heating experiments, typically at 50-70% ^{39}Ar loss. Note that under very slow cooling conditions, different age spectra can be produced from the same sample due to subtle differences in

diffusion domain structure, e.g. sample #70. Sample WN-1 data from C. Roddick, pers. comm., all other age spectra are shown in supplement. **(D-F)** Age spectra and log(r/r_o) plots (second y-axis); note the large change in right y-axis scale between plots. The shape of the log (r/r_o) plot is a function of domain size distribution (relative sizes and volumes) and represents the change in retentivity as domains are progressively outgassed. Cross-correlations between the age and log(r/r_o) data show the high degree of similarity between age and laboratory diffusion spectra, after methods of Lovera et al. (2002). Under monotonic cooling conditions, a high degree of correlation (i.e. >0.90) is expected between $^{40}Ar/^{39}Ar$ age spectra and the laboratory-derived Arrhenius relationship.

Age spectra from the North Caribou Terrane (NCT) at the center of the western Superior craton (sample #70) exhibit ^{39}Ar release patterns that span from >2 Ga to approximately 1.1 Ga, whereas to the southwest, in Manitoba, a Superior Province sample (WN-1) shows Ar release from ca. 1.7 Ga to nearly 1.0 Ga. THO rocks in central Saskatchewan yield age spectra spanning the entire Mesoproterozoic. A sample from the Grenville Province in the Central Metasedimentary Belt (CMB) of Quebec (OL13) yields an age spectrum indicative of initial rapid cooling with slower post-orogenic cooling through the Cambrian. Age spectra from the northern Athabasca Basin margin denote very rapid cooling at ca. 1.7 Ga (fig. 3A; cf. Flowers et al. 2006a,b) and other $^{40}Ar/^{39}Ar$ data from the southern Athabasca basin basement are also in agreement. Despite the small number of samples relative to the spatial distribution, the most striking observation from these data is the consistent and geographically extensive slow cooling signal recorded across the southern craton. Age gradients are large, collectively spanning the majority of the Proterozoic, amounting to ca. 500-600 million years or up to ca. 800-900 million years of cooling in the Superior Province. These gradients are generally consistent with time-integrated slow cooling on the order of <0.5°C/Ma for cratonic rocks during the Proterozoic (an alternative interpretation, involving mild reheating is also possible; see below).

2.3 Thermal history data and modeling

In determining sample-specific kinetic information and the related diffusion-domain information, we used the ranges in K-feldspar kinetics (i.e. diffusivity and activation energies) reported by Lovera et al. (1997) as a cross-check to rule out any unreasonable values. Along with the observed age spectra and their uncertainties, these sample-specific parameters served as the primary input for our thermal history inversions. We performed thermal history modeling only on samples that obey K-feldspar MDD theory, i.e. age spectra consistent with slowly-cooled feldspars should systematically increase in age from 0 to 100% cumulative ^{39}Ar release, or alternatively, individual steps should remain consistent or equal in age during stepwise-degassing of ^{39}Ar (discounting low- or high-temperature excess Ar; fig. 2A-C). Disturbed spectra with intermediate age maxima, saddle-shaped spectra, or high levels of age discordance between individual heating steps were excluded from modeling and interpretation. Age-spectrum steps were modeled only if released below the point of sample breakdown due to partial melting (usually at temperatures of ~1100°C). We used the *Arvert* code for thermal history inversion (www.ees.lehigh.edu/EESdocs/geochron/software.html), which utilizes a Controlled Random Search (CRS) algorithm to search for a family of suitable solutions (see supplementary material for more information). The algorithm searches parameter space for thermal histories that minimize misfit between observed and predicted age spectra and additional mineral ages using the mean square of the weighted deviates (MSWD) as the objective function for assessing the agreement between model predictions and observed data. Explicit high-temperature constraints, variable heating and cooling rates, and additional apatite or zircon (U-Th)/He ages can be included and modeled along with the K-feldspar age information. Thus, below we discuss those samples that are most representative of the regional dataset and are reliable indicators of the overall cooling histories for the southern craton.

2.3.1 Western Superior Province, Ontario

Sample #70 is a tonalite-trondjhemite granite from the Neoarchean North Caribou Terrane of the western Superior Province, one of the oldest Archean cratons in North America (Percival et al., 2006). Most of the rocks in this region record 3.1-2.6 Ga magmatic activity and hydrothermal alteration (Kelly et al., 2017) and a nearby sample yields zircon and titanite U-Pb ages of 2858 ± 5 Ma and 2854 ± 7 Ma, respectively (Van Lankvelt et al., 2016). Ti-in-zircon thermometry indicates zircon crystallization temperatures of 752 ± 37°C (Van Lankvelt et al., 2016). Apatite fission-track (AFT) ages from this region of Ontario are ca. 350-500 Ma (Kohn et al., 2005). A single-grain apatite (U-Th)/He analysis for this sample yields a conventional (U-Th)/He age of 531.5 ± 11.5 Ma (1σ) and a continuous ramped heating (CRH) corrected age of 403 ± 40 Ma (1σ; McDannell et al. 2018). The CRH ^{4}He release spectrum for this sample reveals several discrete He gas spikes throughout the duration of the laboratory heating experiment (see supplement figure). Subsequent ion microprobe analysis on the CAMECA ims-1270 at UCLA and SEM imaging revealed that #70 apatite is characterized by high U-bearing, ~2 μm xenotime inclusions, which likely contribute to anomalous CRH behavior.

Sample WN-1 is an Archean granodiorite from the Winnipeg River Subprovince of the Superior craton from the Lac du Bonnet Batholith emplaced at 2665 ± 20 to 2568 ± 23 Ma (Everitt et al., 1996 and references therein). This sample was from a borehole core at 30.5 m depth and was previously dated via biotite $^{40}Ar/^{39}Ar$, yielding an age of 2291 ± 27 Ma (Hunt and Roddick, 1988). An apatite sample (URL-6-2) at nearly the same depth as WN-1 has an AFT age of 373 ± 29 Ma (Feinstein et al., 2009). These AFT borehole data yield ages of ca. 370 to 250 Ma, decreasing in age with depth.

Our K-feldspar $^{40}Ar/^{39}Ar$ data were modeled assuming monotonic cooling and yielded thermal histories as follows. The majority of samples show cooling rates of ~0.3°C/Ma from ca. 1.9 to 1.5 Ga and relatively slower cooling rates of ≤0.1°C/Ma from ca. 1.5 to ca. 1.1 Ga. Net cooling over this entire interval of 0.8 Ga was only on the order of 100°C (fig. 3). Superior craton $^{40}Ar/^{39}Ar$ spectra from samples #70 and WN-1 in particular display thermal histories indicative of prolonged mid-to-upper crustal residence during the Proterozoic, with time-integrated slow cooling on the order of ~0.1°C/Ma until ca. 1.0 Ga, when marginally faster cooling resumed to bring these rocks to the surface (fig. 3).

It is important to assess the assumption that these samples experienced only monotonic cooling. We use sample #70 to illustrate this non-unique nature of thermal history modeling and establish the resolution and sensitivity of our feldspar thermal histories more generally (fig. 3D). Three monotonic and six non-monotonic thermal-history model inversions were performed under different starting conditions, i.e. permissible cooling rates, number of time nodes, explicit temperature constraints, and achievement of varying degrees of convergence. Both monotonic and reheating histories overlap very well over the portions of the model constrained by observations (fig. 3D), however models that allow reheating indicate that if it did occur, it would have been during the interval 1.45 to 1.1 Ga. High amounts of reheating (>>200°C) erase T-t information through thermal resetting of diffusion domains in K-feldspar, so our data imply that reheating of any amount between ~50°C up to ~200°C is allowed. The best-fitting models favor minimal reheating of ~50°C prior to 1.0 Ga, but, of critical importance still place this sample at temperatures of ~180-200°C near 1.1-1.0 Ga.

2.3.2 Trans-Hudson Orogen, Saskatchewan

The transition from the exposed Precambrian shield bedrock to Phanerozoic cover of the Western Canada basin is exposed in north-central Saskatchewan. Southern Saskatchewan is made up of multiple deformation fronts that welded the ca. 2.4 to 3.1 Ga Sask, Hearne, Rae, and Wyoming cratons during ca. 1.9-1.8 Ga Trans-Hudson orogenesis and suturing of Laurentia (Hoffman, 1988; Bickford et al., 2005). K-feldspar step-heating results from samples of the Athabasca Basin margin in the Chipman Domain basement of the Snowbird Tectonic Zone are tightly constrained by stratigraphy (Rainbird et al., 2007) and existing high-temperature titanite, apatite, and rutile U-Pb ages of ca. 1.90-1.75 Ga, $^{40}Ar/^{39}Ar$ hornblende, muscovite, and biotite ages of 1.78-1.72 Ga (Schneider et al., 2007), and low-temperature zircon and apatite (U-Th)/He ages of 1.70-1.63 Ga and 0.71-0.55 Ga, respectively (Flowers et al., 2006a,b). All of these data define an early rapid cooling event that left Athabasca rocks at or near the surface since ca. 1.6-1.5 Ga.

Our Athabasca K-feldspar data show excellent agreement with existing thermochronologic data (fig. 3). Of most significance, the two K-feldspar samples from the Athabasca Basin margin (samples 02-123A and 02-76B) yield the fast-cooling thermal histories one would predict from existing independent thermochronological results. This is a strong argument for the validity of applying the $^{40}Ar/^{39}Ar$ MDD method to very old rocks, and indicates our other thermal-history results are reliable.

Thermal histories derived from other Athabasca samples are consistent with results obtained from neighboring terranes. Sample SK-KM-01 shows either slow cooling from ca. 1.5 to ca. 1.1 Ga or modest reheating from ca. 1.3 to 1.1 Ga (sample furthest south; fig. 1). Although, the K-feldspar age spectrum from the southern Athabasca basin (SK-KM-03) shows some evidence of low-temperature thermal alteration (Lovera et al., 2002), the modeled T-t history aligns well with the samples from the northern Athabasca basin margin. Whereas the central Saskatchewan sample from the Sask craton (SK-KM-01) exhibits a T-t history similar in style and timing to the Superior Province samples.

2.3.3 Grenville Province, Quebec

Sample OL13 is a syenogranite from the CMB, ~100 km northwest of Ottawa. This area underwent upper-amphibolite facies metamorphism at temperatures and pressures of 700 ± 50°C and 6.5-7.5 kbar (Kretz, 1994) during the Elzivirian and Ottawan phases of the Grenville orogeny. Biotite from this sample yielded a $^{40}Ar/^{39}Ar$ total-gas age of ca. 0.88 Ga and a nearby sample from the same outcrop has a hornblende $^{40}Ar/^{39}Ar$ age of 0.98 Ga (Schneider et al., 2013). Moreover, titanite from hornblende-pyroxene granites in the CMB yield U-Pb ages of ca. 1.02 Ga (Kennedy et al., 2010), allowing the higher temperature Ar systems to constrain ~300-500°C from nearly 1.0 Ga to 0.8 Ga. Collectively these data suggest orogenic cooling of ~2°C/Ma. Our K-feldspar MDD monotonic thermal histories imply rapid cooling through ~250-200°C, followed by post-orogenic slow cooling of ~0.1-0.2°C (fig. 3B). Again, these K-feldspar results are consistent with existing independent geochronological results.

2.4 Summary of results

Like for all thermochronometers, the use of the $^{40}Ar/^{39}Ar$ MDD method to resolve time-temperature histories over protracted periods requires that diffusion mechanisms and kinetics determined in the laboratory over short timescales apply to much longer intervals; i.e., the assumption is that there are no additional slow, low-temperature argon-loss mechanisms that become significant at billion-year timescales (fig. 2D-F). The agreement of our MDD results with independent geochronologic studies of the Athabasca Basin and the Grenville Province localities clearly confirms that the $^{40}Ar/^{39}Ar$ MDD method works and can be reliably applied to very old rocks, providing confidence that other samples are equally informative.

As a result, it is significant that combined $^{40}Ar/^{39}Ar$ MDD K-feldspar data and high- and low-temperature thermochronology reveal much different histories for samples from the southern Canadian Shield (fig. 3B-D). Monotonic cooling-only scenarios exhibit broad agreement between locations in the south-central craton for a period of very slow cooling from ca. 1.5-1.0 Ga, suggesting that these rocks were at elevated temperatures of ~200°C near 1.0 Ga and did not cool quickly to near-surface temperatures. Scenarios allowing reheating indicate that if heating occurred, it was prior to 1.0 Ga, and over a prolonged period coincident with the duration of the Grenvillian orogeny (ca. 200-300 m.y.) and ≈ 50°C to 200°C in magnitude. Nonetheless, both monotonic and reheating histories still suggest rocks were at temperatures near 200°C in the latest Meso- to earliest Neoproterozoic. Moreover, Schwartz and Buchan (1982) examined the thermal overprinting of remnant magnetization on 1.24 Ga Sudbury dikes in southern Ontario. They calculated the country rock was at a paleo-depth of 9.5 ± 2.5 km and temperatures of 267 ± 11°C at the time of dike emplacement, in general agreement with our $^{40}Ar/^{39}Ar$ data.

Rocks at temperatures of 200°C in the late Proterozoic pose an interesting problem. Invoking monotonic cooling and an average cratonic geothermal gradient of 20°C/km, such temperatures would imply that these rocks were in the mid-crust. If reheating occurred via long-term burial, thick sediment packages would be required, and following peak reheating, rocks would then need to be re-exhumed. Therefore, significant burial heating is an unlikely mechanism, although the next sections discuss possible contributing mechanisms to explain these thermal histories.

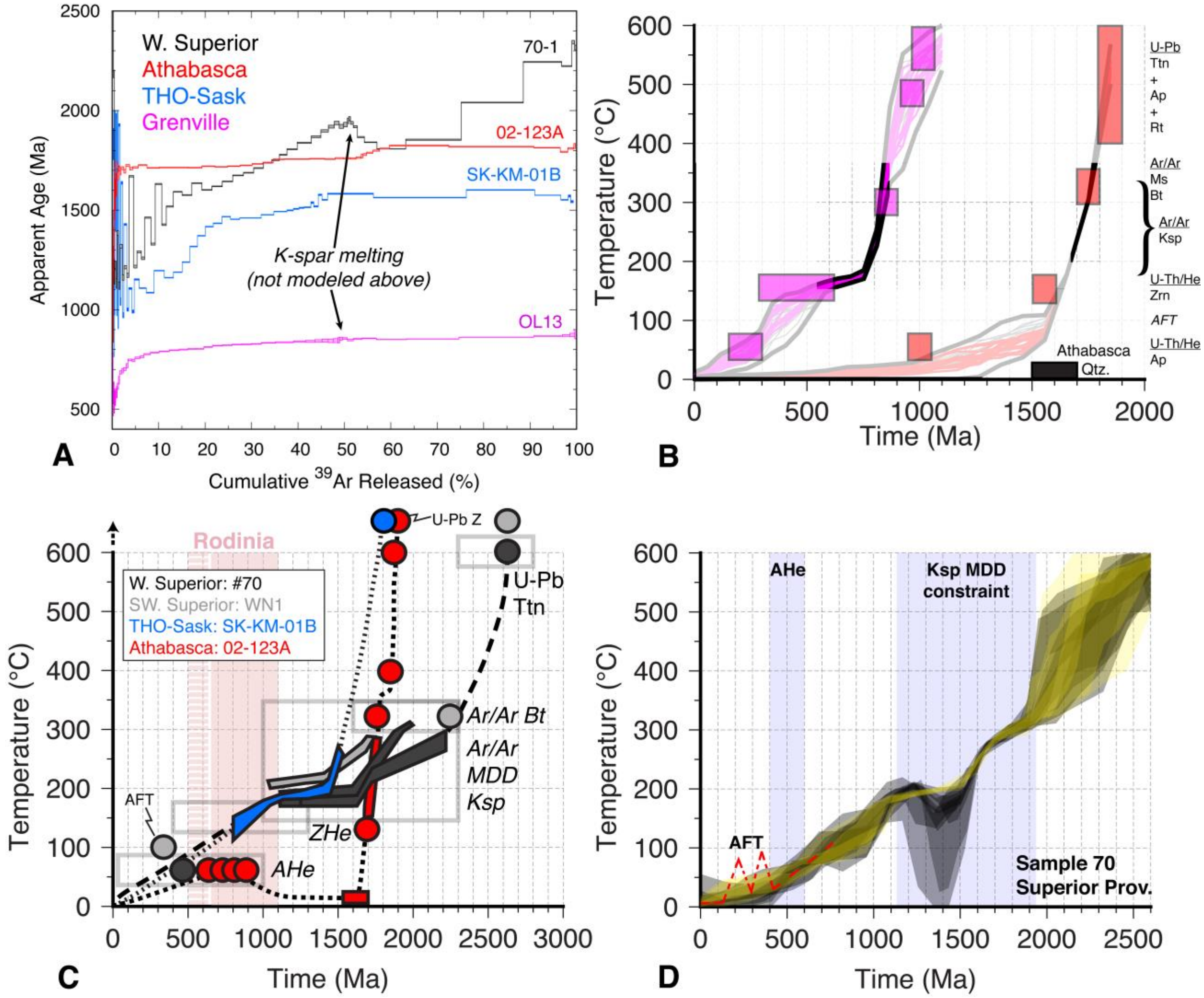

Figure 3. (A) K-feldspar MDD ^{40}Ar/^{39}Ar age spectra from the Western Superior Province (#70; black), Athabasca Basin sample (02-123A; red), THO-Sask craton (SK-KM-01B; blue) and the Grenville Province (OL13; violet). **(B)** Monotonic cooling-only T-t inversions corresponding to colored spectra in fig. 3A (sample 70 T-t monotonic and reheating-allowed results in fig. 3D). These models show excellent agreement with independent higher- and lower-temperature thermochronologic constraints indicated by colored boxes. The MDD spectra can only constrain T-t histories from ~150 to 350°C, illustrated by the T-t envelopes for the 15 best solutions bounded by heavy black lines; steps with excess Ar early in gas release and those after K-feldspar melting at ~1100°C were not modeled. Black box is constraint given by depositional age range of Lake Athabasca sediments. Full age-constraint references are given in supplementary material. **(C)** Summary of constrained T-t envelopes for monotonic-only K-feldspar MDD thermal histories of samples from the Western Superior Province (Samples #70 and WN-1: black and gray symbols), Sask craton sample SK-KM-01B from central Saskatchewan (blue symbols), and a highly constrained Athabasca sample from the Chipman Domain (red symbols); refer to figure 2 for other age spectra that are not shown in fig. 3A. Colored dots correspond to bracketing high and low-temperature data for each sample. Gray boxes show the typical temperature range addressed by other thermochronologic systems. Notice truncated y-axis as U-Pb in zircon closure temperatures are in excess of 850-900°C. Pink band shows the time span of the Rodinia supercontinent from ca. 1.1 Ga to breakup near ca. 0.7-0.6 Ga. Athabasca constraints from Flowers et al. (2006a,b) **(D)** Sample #70 (from NCT,

western Superior Province) thermal-history inversions showing the difference between three monotonic cooling models (yellow envelopes) and six models allowing reheating (black envelopes). Both sets of solutions place these rocks near temperatures of 300°C at ca. 1.8 Ga and >150-200°C at ca. 1.0-1.1 Ga. Regions explicitly constrained by $^{40}Ar/^{39}Ar$ and apatite (U-Th)/He data are shown. Dashed line shows Phanerozoic T-t path from burial/exhumation histories derived from summarized AFT data of Kohn et al. (2005) and Feinstein et al. (2009) during Paleozoic and Mesozoic burial of up to ~1.1-1.2 km from building of the Appalachians. Zrn = zircon, Ttn = titanite, Rt = Rutile; Ms = Muscovite; Bt = biotite; Ksp = K-feldspar; AHe = apatite (U-Th)/He; ZHe = zircon (U-Th)/He; AFT = apatite fission-track

3. Discussion

Samples from the Canadian Shield bedrock at Lake Athabasca and in the Grenville Province document and support the more 'traditional' views of craton stability and rapid post-orogenic decay, respectively, because their thermal histories suggest rapid exhumation to the surface followed by relative stasis. However, unexpectedly some regions appear to have behaved very differently, suggesting more complex or protracted thermal histories and thus more complex lithospheric evolution. Thermal histories from the Archean-Paleoproterozoic southern Canadian Shield (Western Superior Province and Sask craton) indicate rocks there were instead at temperatures near 200°C at 1.0 Ga. No matter how they were achieved, temperatures of 200°C would represent a depth of about 10 km using the maximum geothermal gradient estimate for the Superior Province from Jaupart et al. (2014). This depth and the temperature it is based on provide a serious challenge to conventional views of long-term cratonic stability.

3.1 Burial heating and geotherm variability

One interpretation of our MDD data is that some or all of the 200°C temperatures at 1 Ga were the result of reheating due to sediment burial or changes in heat flow. To achieve 180-200°C of burial reheating, significant accommodation space and sediment source would have been required. From ca. 1.3-1.0 Ga the Superior Province witnessed deformation from Grenvillian contraction and foreland sedimentation on one side and MCR extension on the other. If one attributes all or some of the reheating to burial, an explanation would be required as to how such a horizontally and vertically extensive package of sediments could be deposited at 1 Ga, far inboard of the Grenville orogen. If this were the case, there would remain a reasonably thick rock column to be removed afterwards, since our samples were collected from the current bedrock surface. It has been recognized, however, that the central craton has undergone platform sedimentation numerous times since ca. 2 Ga and the spatially-dispersed sediments found across the landscape suggest that most of Laurentia was cyclically buried by sediment in the past (Davidson, 2008). Detrital zircon provenance studies and paleo-current markers also reveal that large continental-scale fluvial networks transported sediments shed from the Trans-Hudson orogen (Rainbird et al., 2007) and later high topography of the Grenville orogen (Rainbird et al., 1992). These extensive fluvial systems locally buried the craton in a veneer of sediment up to 1 km thick. However, it is difficult to assert that the majority of the southeastern Canadian Shield experienced sheet-like deposition thick enough to appreciably reheat K-feldspar MDD samples.

To test this, we used a simple model of a steady-state crustal geothermal gradient to estimate the effects of burial heating resulting from 5 km of sediments overlying a 10 km thick radiogenic upper crust. The model incorporated moderate bulk crustal heat production (A_1) of 3 $\mu W/m^3$, basin heat production (A_2) of 5 $\mu W/m^3$, surface temperature of 10°C, basal heat flow of 30 mW/m^2, and assumed uniform thermal conductivity (K) of 2.5 W/(m·K). The base of the basin sediments would reach temperatures of 125°C; therefore, an intracratonic basin would have to be greater than 5 km thick to achieve temperatures near those experienced by our rock samples, which seems unreasonable for this setting. Heating to 200°C at 5 km depth could be achieved through a combination of higher basin heat production, a thicker radiogenic upper crust, or greater basal heat flow. This end-member burial scenario is possible but improbable. Moreover, there is still no evidence of any proximal intracratonic basins of appropriate Mesoproterozoic age, and most of the nearby preserved sediments were shed prior to and after Laurentian assembly at ca. 1.8 Ga (Davidson, 2008).

It seems unlikely, at least under current views about cratonic evolution, that 200°C temperatures at shallow depths could have been related to an abnormal geothermal gradient, and not burial, because thermal gradients approaching 40-50°C or greater would have been required and sustained on the order of a hundred million years or more to match the thermal-history model results. Most geothermal perturbations are short-lived. An estimate of the Paleoproterozoic steady-state geotherm of the Witwatersrand Basin is 27°C/km with an increase to ~40°C/km during ~2 Ga Bushveld magmatism and mafic underplating that lasted for <30 m.y. before returning to steady-state conditions (Frimmel, 1997). At the scale of the lithosphere, plume-head thermal diffusion effects have been modeled for the Great Meteor Hotspot of the northeastern

North American craton, and a 500°C anomaly initially confined to the lower 50 km of the cratonic lithosphere decays to 80°C at 200 km depth over 120 m.y., while the middle to upper crust records very little perturbation (Eaton and Frederiksen, 2007). Therefore, it is reasonable to believe that thermal diffusivity through thick cratonic lithosphere is long (ca. 10^9 years) and the amount of basal heat required to substantially warm the middle and upper crust is high.

3.2 Erosional exhumation via isostasy

Whether there was reheating or not, it is not feasible to obtain 10 km of basement exhumation through simple erosional isostatic return over the last 1 billion years because reasonable values of upper/lower crust and mantle densities (e.g. Schmandt et al., 2015) would require a crustal thickness of over 90 km at 1 Ga to permit this much total exhumation. Erosion of this amount is also conceptually difficult, given the widely assumed low continental freeboard that one might infer from such slow cooling between 2 Ga and 1 Ga. The "constant freeboard model" asserts that under isostatic-equilibrium conditions, the denudation of exposed cratonic terranes and accumulation of sediments on low-lying craton margins leads to continental elevations trending towards equilibrium with mean sea level over long timescales (Eriksson et al., 2006). This carries implications for the erosional potential at 1 Ga, where any significant cratonic topography would have been substantially eroded and the continental interior would have to be above sea level for excess elevation to be isostatically balanced by crustal thickness.

There are also discrepancies regarding total erosion levels and crustal thickness approximations. Abbott et al. (2000) estimate that the worldwide total crustal erosion that occurred in Archean and early Proterozoic terranes was minimal, ranging from 2.5-7.8 km (4.1 km avg.) and 1.65-4.4 km (2.5 km avg.), respectively. Pressure-temperature estimates of rocks exposed at the surface (fig. 4) also demonstrate that large portions of the western Superior Province have been exhumed a *total* of 10-15 km (~4-6 kbar), whereas present Moho depths are at ~40 km, which places limits on maximum crustal thicknesses on the order of ~45-60 km (Percival et al., 2006; Percival et al., 2012). Integrated geothermal constraints and prior P-T studies suggest that exposed North Caribou terrane metamorphic rocks in the Western Superior Province were eroded ~11 km total (Jaupart et al., 2014). With the known erosion levels and estimated original crustal thickness, exhumation-driven cooling from ~200°C at 1.0 Ga would require post-Archean crustal thickening, with the majority of total cratonic exhumation being delayed until the late Proterozoic. In short, petrologic observations combined with our K-feldspar MDD data would suggest that exhumation of virtually all of the southern Canadian Shield was deferred to (and after) ca. 1 Ga.

3.3 Magmatic underplating

Crustal thickening by underplating can reconcile the renewed cooling seen in our thermal histories with reasonable levels of crustal erosion. Mafic underplating is integral to many models of Archean-Proterozoic geodynamics and crustal growth, where high-density material is added to the base of the crust (Thybo and Artemieva, 2013). The emplacement of 10-25 km of underplated material can rapidly thicken and add significant crustal volume under normal lithospheric thermal conditions (Cox, 1993; Furlong and Fountain, 1986). Adiabatic decompression of the mantle forms melt that typically has a density of 3000-3100 kg/m^3, which lies between typical mantle and crustal values, therefore increasing the potential for mafic-melt trapping in the lower crust that displaces denser lithospheric mantle thus producing surface uplift (Allen and Allen, 2013). The high seismic velocities observed in the lower crust of underplated regions seems to verify this (Thybo and Artemieva, 2013), and underplating has been proposed for other cratons such as the Kaapvaal craton in South Africa that experienced crustal inflation and uplift during Karoo province magmatism (Cox, 1993).

3.3.1 Underplating at the Mid-Continent Rift

The Mid-Continental Rift affected the Superior craton at ca. 1.1 Ga and various models exist for initiation, either a hot mantle plume that impinged on the lower lithosphere (e.g. Nicholson and Shirey, 1990) or as a far-field extensional response to compressional stresses during the Grenville orogeny (e.g. Whitmeyer and Karlstrom, 2007), or a combination of the two (Stein et al., 2015). Northward propagation of the MCR was inhibited by the strong Superior lithosphere, and the shape and position of the rift was influenced by the stable craton margin (Ola et al., 2016). There is evidence for basaltic underplating of the Archean lower crust during the 1.1 Ga failed rifting of the mid-continent (e.g. Behrendt et al., 1990; Tréhu et al., 1991) with a 10-20 km thick layer between the lower crust and upper mantle interpreted as underplated magmatic material (Zhang et al., 2016). Other seismic studies suggest that the crust is unusually thick at the

MCR from reactivation and underplating during late stages of the Grenville orogeny (e.g. Shen et al., 2013). Geophysical modeling also shows a similar Moho density contrast interpreted as thickened crust due to mafic underplating that is >15 km thick at the base of the southwestern Superior crust that occurred during Trans-Hudson (1.8 Ga) suturing of the Yavapai-Superior-Wyoming blocks (Thurner et al., 2015).

3.3.2 Underplating in the Superior Province

In general, Proterozoic crust is consistently thicker (40-55 km) and is characterized by greater lower crustal seismic velocities, when compared to Archean terranes (Durrheim and Mooney, 1991). The North Caribou Terrane is the core of the western Superior Province and is contrarily characterized by thicker crust (40-43 km), given its Archean affinity (Percival et al., 2012; Perry et al., 2002). There is also regional evidence for thickened crust, with thickening from 40 km in the NCT to 45 km towards Lake Superior at the locus of MCR rifting. This is markedly different from the eastern Superior craton, which is ~35 km thick on average (Perry et al., 2002). The greater crustal thickness in the western Superior compared to the eastern Superior and thickening towards the MCR supports the compiled worldwide observations for post-Archean underplating and crustal thickening to first-order (Durrheim and Mooney, 1991; Thybo and Artemieva, 2013).

The lower crust-upper mantle anomaly that exists in the western Superior Province has been interpreted to be the result of Archean imbricate oceanic slab accretion (in the Wabigoon Subprovince) under the NCT during lithospheric assembly that appears as Moho-penetrating high-velocity regions in seismic profiles (Musacchio et al., 2004; White et al., 2003; fig. 4). Nitescu et al. (2006) hypothesize that the oceanic slab argument is not conclusive and that continuity in lower crustal densities align more with a model of post-accretion thermal softening and lower crustal reworking. They argue that the mass anomalies within the lower crust of the Wabigoon Province and those terranes to the south instead represent dense underplated, anisotropic mafic rocks from MCR extension and magmatism.

Thermobarometry and the compositional differences with depth (fertile vs. depleted) found between garnets in the Proterozoic Kyle Lake and Paleozoic-Mesozoic Attawapiskat kimberlites (just east of Kyle Lake; fig. 4) in the western Superior Province signify a modified mantle composition or reworked structure between parts of the western province due to extension and basalt emplacement (Scully et al., 2004). The thickened crust of the rift flanks experienced uplift due to increased buoyancy and the "topographic doming" model of Allen et al. (1992) proposes MCR melting from a mantle plume source affected a large portion of the western Superior lower crust with underplating occurring up to 650 km from the rift itself (fig. 4). This model is reinforced by the aforementioned high-density, lower-crustal layer thickening towards the MCR and the presence of synchronous Keweenawan dikes (and kimberlites) north of Lake Superior (Ernst and Bleeker, 2010), which are typically believed to be the surface expression of underplated mafic rocks (Nelson, 1991).

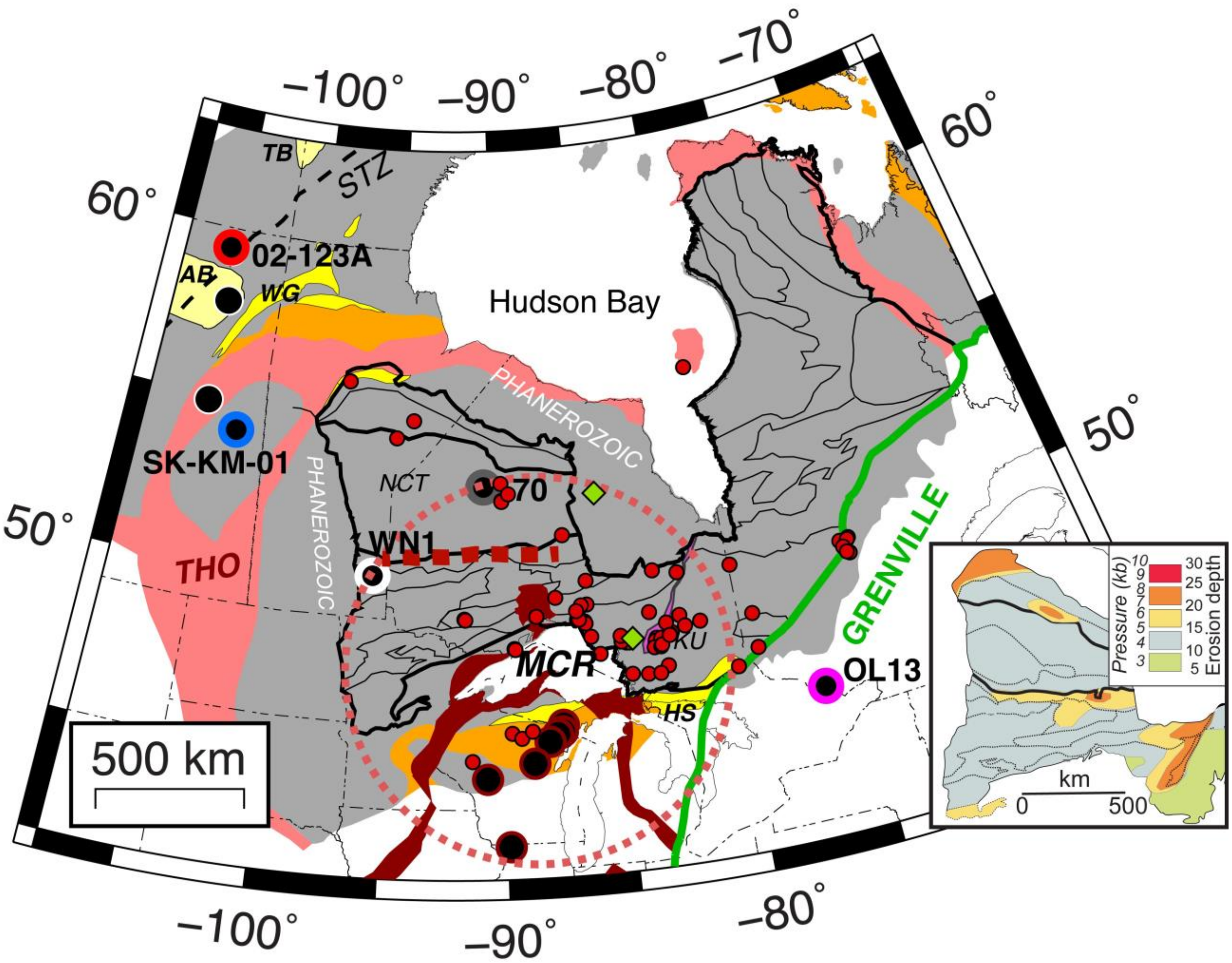

Figure 4. Map of the southern Canadian Shield modified from Whitmeyer and Karlstrom (2007) with geologic sub-terranes of the Superior Province outlined in black from Percival et al. (2006). Inset map shows erosion depths in km and P-T estimates across the western Superior Province, adapted from Percival et al. (2012). K-feldspar MDD samples dated in this study via $^{40}Ar/^{39}Ar$ are shown by black dots and the MCR likely disturbed those with dark red outlines (see supplement for age spectra). Small red dots are from the Geological Survey of Canada geochronology database filtered by 0.95-1.2 Ga date

range, largely representing mafic dike ages using various dating methods. Note: the red points in NW Superior are likely related to the ca. 1.26 Mackenzie dike swarm originating in the Slave craton up to 3000 km to the northwest. The dark red dashed line near the southern NCT boundary is the approximate high seismic velocity region from White et al. (2003); Musacchio et al. (2004) and gravity anomalies of Nitescu et al. (2006) interpreted as a result of MCR underplating. Red dashed circle is the region that could theoretically be affected by topographic doming up to 650 km away from the MCR nucleus, from Allen (1992). Light green diamonds are Kyle Lake (north) and Abitibi kimberlites (south) of ca. 1.1 Ga age. All other map elements are referenced in figure 1.

The western Superior Province shows evidence of significant deformation related to dike swarms and rifting at ca. 1.27 Ga from the Mackenzie plume and dike system (in the Slave Province far to the northwest) and the ca. 1.1 Ga MCR mantle plume (fig. 4). This later magmatism produced a >15 km underplated, thickened crust at the MCR, which tapers to approximately ≤10 km thickness to the north in the western Superior Province, based on the aforementioned studies showing a high velocity lower crustal layer (e.g. Musacchio et al., 2004; Nitescu et al., 2006). Lower-crustal magmatic underplating causes thickening and drives isostatic compensation to account for added mass. To model the appropriate crustal response, we used typical crustal densities, as well as those from Thurner et al. (2015) of 2950, 3100, and 3300 kg/m^3 for the crust, underplate material, and mantle, respectively, to simulate isostatic uplift and erosion (table 1). Using equations from Allen and Allen (2013), under normal lithospheric conditions with the surface at sea level, the pressure at depth, y_L at the base of the lithosphere is:

$$y_c \rho_c g + (y_L - y_c) \rho_m g \tag{1}$$

Where g is gravitational acceleration, Y_c and Y_L are the thickness of the crust and lithosphere and ρ_c and ρ_m are the crustal and mantle densities, respectively. The pressure at the same depth, under uplifted lithosphere with an underplate (H) is given as:

$$y_c \rho_c g + H \rho_h g + (y_L - y_c - H + U) \rho_m g \tag{2}$$

Equation (3) describes the uplift (U) relationship under Airy isostasy conditions and the balance of lithospheric columns at the base of the lithosphere (compensation depth); where H is the underplate thickness, ρ_h is the density of the underplate, and ρ_m is the density of the mantle.

(3)

$$U = H\left(1 - \frac{\rho_h}{\rho_m}\right)$$

Erosional denudation (D) is included, where ρ_c is the crustal density:

$$D = H\left(\frac{\rho_m - \rho_h}{\rho_m - \rho_c}\right) \quad (4)$$

Underplate thicknesses of 3 to 10 km (table 1) produce respective isostatic uplift amounts of ~200 to ~900 m, resulting in ~1700 to 6000 m of denudation (assuming no change in surface elevation).

Table 1: Isostatic uplift and denudation for magmatic underplating

Underplate, H (m)	ρ_c (kg m^{-3})	ρ_h (kg m^{-3})	ρ_m (kg m^{-3})	Uplift, U (m)	Denudation, D (m)
3000	2800	3000	3300	273	1800
5000	2800	3000	3300	455	3000
10000	**2800**	**3000**	**3300**	**909**	**6000**
15000	2800	3000	3300	1364	9000
20000	2800	3000	3300	1818	12000
3000	2950	3100	3300	182	1714
5000	2950	3100	3300	303	2857
10000	2950	3100	3300	606	5714
15000	2950	3100	3300	909	8571
20000	2950	3100	3300	1212	11429

Table 1. The role of magmatic underplating of variable thickness and density is shown. The relation of crustal density changes inducing an isostatic response causing uplift and corresponding erosion at the surface. Mafic underplating of typical thicknesses for the crust between 3-20 km (variable density) can cause up to ~275-1800 m of uplift, resulting in ~2-12 km of erosion. Note: multiplying the underplate thickness by a factor of 0.6 yields the amount of denudation for typical crustal/underplate/mantle densities of 2800/3000/3300 kg/m^3. The 10 km underplate thickness in bold is a likely scenario for the western Superior Province from geophysical and gravity data interpretations.

3.4 Summary of mechanisms to explain T-t histories

Our key finding is that $^{40}Ar/^{39}Ar$ K-feldspar MDD data establish that rocks in the southern Canadian Shield were at ~200°C at ca. 1 Ga. Our preferred mechanism to provide the required exhumation required to expose these rocks is magmatic underplating, which is a feasible way to thicken the crust in the Proterozoic and cause isostatic disequilibrium that would necessitate erosional re-equilibration. Assuming a 20°C/km geothermal gradient, monotonic cooling-only thermal histories for the western Superior Province indicate that the surface rocks from which our feldspar samples were taken were at depths of up to 10 km in the crust at ca. 1.0 Ga. Extreme crustal thicknesses of ~90 km would be required if rocks were simply isostatically exhumed over the past 1 billion years. Our data imply that if rocks resided in the middle crust, then the majority of craton unroofing was delayed until Rodinian time near the end of the 'Boring Billion.' Alternatively, reheating between ca. 1.45-1.1 Ga on the order of ~50-200°C is allowed by the $^{40}Ar/^{39}Ar$ MDD thermal histories, depending on the amount of pre-heating exhumation. Our best thermal history models favor the lower end of this reheating range near ~50°C. Reheating of this magnitude could be achieved through < 3 km of sedimentary burial or significant upper crustal heating (having no obvious source) prior to 1.0 Ga. Ruling out a purely thermal anomaly in the shallow crust, burial of partially exhumed exhumed metamorphic rocks still leaves the samples at depth at 1.0 Ga. However, monotonic thermal histories seem preferable, as there is no sedimentary evidence to support moderate, widespread burial in many areas, nor evidence of a sustained elevated crustal geotherm that would have to be maintained over tens to hundreds of millions of years. The time span of the sub-Cambrian Great Unconformity suggests that erosion was the more dominant process across the exposed Laurentian shield at the end of the Proterozoic.

4. Conclusions

Our $^{40}Ar/^{39}Ar$ MDD K-feldspar data fill a large observational gap and yield thermal histories entirely consistent with other geochronologic and geologic information available for younger cratonic terranes: Athabasca Basin and Grenville rocks show thermal histories of rapid cooling and exhumation, in agreement with sedimentary data and high-temperature geochronology – and our K-feldspar age spectra lend further support for this observation. $^{40}Ar/^{39}Ar$ age spectra and thermal-history solutions from southern craton localities suggest very slow cooling of <0.5°C/Ma throughout the Earth's middle age, whereas histories allowing reheating permit some minor burial prior to ca. 1.0 Ga. Despite these subtle differences, the most important point is that regardless of the type of history, southern craton rocks that are now at the surface were at temperatures of ~200°C in the latest Proterozoic.

The known levels of total erosion and P-T constraints on the total crustal thickness in the western Superior Province put limits on the regional exhumation history. Proterozoic slow cooling requires crustal thickening on the order of ~10 km near 1.0 Ga to permit the necessary uplift and erosion to explain the thermal histories derived from $^{40}Ar/^{39}Ar$ data. We argue that regional magmatic underplating due to ca. 1.1 Ga Mid-Continent rifting thickened the crust and caused exhumation beginning in the very latest Meso- to early Neoproterozoic. This is supported by regional geophysical data that reveals underplating and crustal thickening adjacent to the Mid-Continental Rift in the western Superior Province. Ultimately, there is an issue with timing regarding the amount of exhumation ± burial prior to 1.0 Ga and the estimate of the actual crustal depth of our samples at 1.0 Ga. Our preferred scenario is that after Archean-Paleoproterozoic tectonism, there was ~3-5 km of exhumation in (linked) combination with allowing up to ~1-3 km of Mesoproterozoic burial, followed by ~6-9 km of exhumation starting at ca. 1.0 Ga. Integrated thermochronology and geologic records thus suggest that most cratonic exhumation was delayed until the early Neoproterozoic, during this previously unrecognized exhumation event. Large-scale denudation contributed to the development of the Great Unconformity in North America, preceding the recognized epeirogenic burial and exhumation episodes occurring during the early Phanerozoic.

Acknowledgements

We thank An Yin for editorial handling and T. Mark Harrison and an anonymous reviewer for efficient and insightful reviews that helped clarify the manuscript. We would also like to thank Rob Rainbird at the Geological Survey of Canada, Ottawa for offering additional comments on the manuscript. We also thank Rebecca Flowers at the University of Colorado, Boulder for providing Athabasca samples and Ryan Morelli at the Saskatchewan Geologic Survey for access to granite cores. Thanks to Nancy Joyce of the GSC-Ottawa geochronology lab for finding information about a long-lost sample and thanks to Bruce Idleman for assistance with sample preparation, irradiation logistics, and in running samples at the Lehigh noble-gas laboratory. Natural Resources Canada, Land and Minerals sector, Geological Survey of Canada: PPI contribution ID# 36455.

References

Abbott, D., Sparks, D., Herzberg, C., Mooney, W., Nikishin, A., and Zhang, Y. S., 2000, Quantifying Precambrian crustal extraction: the root is the answer: Tectonophysics, v. 322, no. 1, p. 163-190.

Alexandre, P., Kyser, K., Thomas, D., Polito, P., and Marlat, J., 2009, Geochronology of unconformity-related uranium deposits in the Athabasca Basin, Saskatchewan, Canada and their integration in the evolution of the basin: Mineralium Deposita, v. 44, no. 1, p. 41-59.

Allen, D. J., Hinze, W. J., and Cannon, W. F., 1992, Drainage, topographic, and gravity anomalies in the Lake Superior region: Evidence for a 1100 Ma mantle plume: Geophysical Research Letters, v. 19, no. 21, p. 2119-2122.

Allen, P. A., and Allen, J. R., 2013, Basin analysis: Principles and application to petroleum play assessment, John Wiley & Sons.

Ault, A. K., Flowers, R. M., and Bowring, S. A., 2009, Phanerozoic burial and unroofing history of the western Slave craton and Wopmay orogen from apatite (U–Th)/He thermochronometry: Earth and Planetary Science Letters, v. 284, no. 1-2, p. 1-11.

Behrendt, J. C., Hutchinson, D., Lee, M., Thornber, C., Trehu, A., Cannon, W., and Green, A., 1990, GLIMPCE seismic reflection evidence of deep-crustal and upper-mantle intrusions and magmatic underplating associated with the Midcontinent Rift system of North America: Tectonophysics, v. 173, no. 1-4, p. 595-615.

Bickford, M., Mock, T., Steinhart Iii, W., Collerson, K., and Lewry, J., 2005, Origin of the Archean Sask craton and its extent within the Trans-Hudson orogen: evidence from Pb and Nd isotopic compositions of basement rocks and post-orogenic intrusions: Canadian Journal of Earth Sciences, v. 42, no. 4, p. 659-684.

Cawood, P. A., and Hawkesworth, C. J., 2014, Earth's middle age: Geology, v. 42, no. 6, p. 503-506.

Cox, K., 1993, Continental magmatic underplating: Philosophical Transactions of the Royal Society of London A: Mathematical, Physical and Engineering Sciences, v. 342, no. 1663, p. 155-166.

Davidson, A., 2008, Late Paleoproterozoic to mid-Neoproterozoic history of northern Laurentia: an overview of central Rodinia: Precambrian Research, v. 160, no. 1, p. 5-22.

Durrheim, R. J., and Mooney, W. D., 1991, Archean and Proterozoic crustal evolution: Evidence from crustal seismology: Geology, v. 19, no. 6, p. 606-609.

Eaton, D. W., and Frederiksen, A., 2007, Seismic evidence for convection-driven motion of the North American plate: Nature, v. 446, no. 7134, p. 428-431.

Eriksson, P. G., Mazumder, R., Catuneanu, O., Bumby, A. J., and Ilondo, B. O., 2006, Precambrian continental freeboard and geological evolution: a time perspective: Earth-Science Reviews, v. 79, no. 3, p. 165-204.

Ernst, R., and Bleeker, W., 2010, Large igneous provinces (LIPs), giant dyke swarms, and mantle plumes: significance for breakup events within Canada and adjacent regions from 2.5 Ga to the Present: Canadian Journal of Earth Sciences, v. 47, no. 5, p. 695-739.

Everitt, R., McMurry, J., Brown, A., and Davison, C., 1996, Geology of the Lac Du Bonnet batholith, inside and out: AECL's underground research laboratory, southeastern Manitoba, *in* Proceedings Field excursion B-5: guidebook. Geological Association of Canada–Mineralogical Association of Canada Joint Annual Meeting.

Feinstein, S., Kohn, B., Osadetz, K., Everitt, R., and O'Sullivan, P., 2009, Variable Phanerozoic thermal history in the Southern Canadian Shield: Evidence from an apatite fission track profile at the Underground Research Laboratory (URL), Manitoba: Tectonophysics, v. 475, no. 1, p. 190-199.

Flowers, R., Bowring, S., Mahan, K., Williams, M., and Williams, I., 2008, Stabilization and reactivation of cratonic lithosphere from the lower crustal record in the western Canadian shield: Contributions to Mineralogy and Petrology, v. 156, no. 4, p. 529-549.

Flowers, R. M., Bowring, S. A., and Reiners, P. W., 2006a, Low long-term erosion rates and extreme continental stability documented by ancient (U-Th)/He dates: Geology, v. 34, no. 11, p. 925-928.

Flowers, R., Mahan, K., Bowring, S., Williams, M., Pringle, M., and Hodges, K., 2006b, Multistage exhumation and juxtaposition of lower continental crust in the western Canadian Shield: Linking high-resolution U-Pb and 40Ar/39Ar thermochronometry with pressure-temperature-deformation paths: Tectonics, v. 25, no. 4.

Flowers, R. M., Ault, A. K., Kelley, S. A., Zhang, N., and Zhong, S., 2012, Epeirogeny or eustasy? Paleozoic-Mesozoic vertical motion of the North American continental interior from thermochronometry and implications for mantle dynamics: Earth and Planetary Science Letters, v. 317-318, no. 0, p. 436-445.

Forte, A. M., Moucha, R., Simmons, N. A., Grand, S. P., and Mitrovica, J. X., 2010, Deep-mantle contributions to the surface dynamics of the North American continent: Tectonophysics, v. 481, p. 3-15.

Frimmel, H. E., 1997, Chlorite thermometry in the Witwatersrand Basin: constraints on the Paleoproterozoic geotherm in the Kaapvaal Craton, South Africa: The Journal of Geology, v. 105, no. 5, p. 601-616.

Furlong, K. P., and Fountain, D. M., 1986, Continental crustal underplating: Thermal considerations and seismic-petrologic consequences: Journal of Geophysical Research: Solid Earth, v. 91, no. B8, p. 8285-8294.

Hoffman, P. F., 1988, United Plates of America, the birth of a craton: Early Proterozoic assembly and growth of Laurentia: Annual Review of Earth and Planetary Sciences, v. 16, p. 543-603.

Hunt, P. A., and Roddick, J. C., 1988, A compilation of K-Ar ages: Report 18; Radiogenic Age and Isotopic Studies: Report 2: Geological Survey of Canada.

Jaupart, C., J. C. Mareschal, H. Bouquerel, and C. Phaneuf, 2014, The building and stabilization of an Archean Craton in the Superior Province, Canada, from a heat flow perspective, Journal of Geophysical Research: Solid Earth.

Jordan, T. H., 1978, Composition and development of the continental tectosphere: Nature, v. 274, no. 5671, p. 544-548.

Karlstrom, K. E., and Timmons, J. M., 2012, Many unconformities make one Great Unconformity: Spec. Pap., Geol. Soc. Am, v. 489, p. 73-79.

Kelly, C.J., Schneider, D.A., Jackson, S.E., Kalbfleisch, T. and McFarlane, C.R., 2017, Insights into low-to moderate-temperature recrystallization of zircon: Unpolished crystal depth profile techniques and geochemical mapping. Chemical Geology, 449, p. 82-98.

Kennedy, A.K., Kamo, S.L., Nasdala, L. and Timms, N.E., 2010, Grenville skarn titanite: potential reference material for SIMS U–Th–Pb analysis. The Canadian Mineralogist, v. 48(6), p. 1423-1443.

Kohn, B.P., Gleadow, A.J., Brown, R.W., Gallagher, K., Lorencak, M., Noble, W.P., 2005, Visualizing thermotectonic and denudation histories using apatite fission track thermochronology, in: Reiners, P.W., Ehlers, T.A. (Eds.), Reviews in Mineralogy and Geochemistry. Mineralogical Society of America Geochemical Society, Chantilly, Virginia, pp. 527-565.

Kretz, R., 1994, Petrology of veined gneisses of the Otter complex, southern Grenville Province: Canadian Journal of Earth Sciences, v. 31, no. 5, p. 835-851.

Li, Z.-X., Bogdanova, S., Collins, A., Davidson, A., De Waele, B., Ernst, R., Fitzsimons, I., Fuck, R., Gladkochub, D., and Jacobs, J., 2008, Assembly, configuration, and break-up history of Rodinia: a synthesis: Precambrian research, v. 160, no. 1, p. 179-210.

Lovera, O. M., M. Grove, and T. M. Harrison, 2002, Systematic analysis of K-feldspar 40Ar/39Ar step heating results II: relevance of laboratory argon diffusion properties to nature, Geochimica et Cosmochimica Acta, v. 66 no. 7, p. 1237-1255.

Lovera, O. M., Grove, M., Harrison, T. M., & Mahon, K. I., 1997, Systematic analysis of K-feldspar 40Ar39Ar step heating results: I. Significance of activation energy determinations. Geochimica et Cosmochimica Acta, 61(15), 3171-3192.

Lovera, O. M., Richter, F. M., and Harrison, T. M., 1989, The 40Ar/39Ar Thermochronometry for Slowly Cooled Samples Having a Distribution of Diffusion Domain Sizes: Journal of Geophysical Research, v. 94, no. B12, p. 17917-17935.

Lovera, O. M., Richter, F. M., and Harrison, T. M., 1991, Diffusion Domains Determined by 39Ar Released During Step Heating: Journal of Geophysical Research, v. 96, no. B2, p. 2057-2069.

McDannell, K. T., P. K. Zeitler, D. G. Janes, B. D. Idleman, and A. K. Fayon, 2018, Screening apatites for (U-Th)/He thermochronometry via continuous ramped heating: Helium age components and implications for age dispersion: Geochimica et Cosmochimica Acta, v. 223, p. 90-106. doi: 10.1016/j.gca.2017.11.031

McDougall, I., and Harrison, T. M., 1999, Geochronology and Thermochronology by the 40Ar/39Ar Method, Oxford University Press, USA.

Mitrovica, J. X., Beaumont, C., and Jarvis, G. T., 1989, Tilting of continental interiors by the dynamical effects of subduction: Tectonics, v. 8, no. 5, p. 1079-1094.

Musacchio, G., White, D., Asudeh, I., and Thomson, C., 2004, Lithospheric structure and composition of the Archean western Superior Province from seismic refraction/wide-angle reflection and gravity modeling: Journal of Geophysical Research: Solid Earth, v. 109, no. B3.

Nelson, K. (1991) A unified view of craton evolution motivated by recent deep seismic reflection and refraction results. Geophysical Journal International 105, 25-35.

Nicholson, S. W., and Shirey, S. B., 1990, Midcontinent rift volcanism in the Lake Superior region: Sr, Nd, and Pb isotopic evidence for a mantle plume origin: Journal of Geophysical Research: Solid Earth, v. 95, no. B7, p. 10851-10868.

Nitescu, B., Cruden, A., and Bailey, R., 2006, Crustal structure and implications for the tectonic evolution of the Archean Western Superior craton from forward and inverse gravity modeling: Tectonics, v. 25, no. 1.

Ola, O., Frederiksen, A., Bollmann, T., van der Lee, S., Darbyshire, F., Wolin, E., Revenaugh, J., Stein, C., Stein, S., and Wysession, M., 2016, Anisotropic zonation in the lithosphere of Central North America: Influence of a strong cratonic lithosphere on the Mid-Continent Rift: Tectonophysics, v. 683, p. 367-381.

Percival, J., Sanborn-Barrie, M., Skulski, T., Stott, G., Helmstaedt, H., and White, D., 2006, Tectonic evolution of the western Superior Province from NATMAP and Lithoprobe studies: Canadian Journal of Earth Sciences, v. 43, no. 7, p. 1085-1117.

Percival, J., Skulski, T., Sanborn-Barrie, M., Stott, G., Leclair, A., Corkery, M., and Boily, M., 2012, Geology and tectonic evolution of the Superior Province, Canada: Tectonic Styles in Canada: The Lithoprobe Perspective. Geol Assoc Canada Spec Paper, v. 49, p. 321-378.

Percival, J. A., and West, G. F., 1994, The Kapuskasing uplift: a geological and geophysical synthesis. Canadian Journal of Earth Sciences, v. 31(7), p. 1256-1286.

Perry, H. K. C., Eaton, D. W. S., and Forte, A. M., 2002, LITH5.0: a revised crustal model for Canada based on Lithoprobe results: Geophys. Jour. Int., v. 150, p. 285-294.

Pollack, H. N., 1986, Cratonization and thermal evolution of the mantle: Earth and Planetary Science Letters, v. 80, no. 1, p. 175-182.

Rainbird, R. H., Heaman, L. M., and Young, G., 1992, Sampling Laurentia: Detrital zircon geochronology offers evidence for an extensive Neoproterozoic river system originating from the Grenville orogen: Geology, v. 20, no. 4, p. 351-354.

Rainbird, R., Stern, R., Rayner, N., and Jefferson, C., 2007, Age, provenance, and regional correlation of the Athabasca Group, Saskatchewan and Alberta, constrained by igneous and detrital zircon geochronology: Bulletin - Geological Survey of Canada, v. 588, p. 193.

Rogers, J. J., and Santosh, M., 2002, Configuration of Columbia, a Mesoproterozoic supercontinent: Gondwana Research, v. 5, no. 1, p. 5-22.

Rudnick, R. L., and Nyblade, A. A., 1999, The thickness and heat production of Archean lithosphere: constraints from xenolith thermobarometry and surface heat flow: Mantle petrology: Field observations and high pressure experimentation: A tribute to Francis R.(Joe) Boyd, v. 6, p. 3-12.

Schmandt, B., Lin, F. C., and Karlstrom, K. E., 2015, Distinct crustal isostasy trends east and west of the Rocky Mountain Front: Geophysical Research Letters.

Schneider, D., Cope, N., and Holm, D., 2013, Thermochronology of the Mont Laurier terrane, southern Canadian Grenville Province, and its bearing on defining orogenic architecture: Precambrian Research, v. 226, p. 43-58.

Schneider, D. A., Heizler, M. T., Bickford, M. E., Wortman, G. L., Condie, K. C., and Perilli, S., 2007, Timing constraints of orogeny to cratonization: Thermochronology of the Paleoproterozoic Trans-Hudson orogen, Manitoba and Saskatchewan, Canada: Precambrian Research, v. 153, p. 65-95.

Schwartz, E.J. and Buchan, K.L., 1982, Uplift deduced from remnant magnetization: Sudbury area since 1250 Ma ago: Earth and Planetary Science Letters, v. 58, p. 65-74.

Scully, K. R., Canil, D., and Schulze, D. J., 2004, The lithospheric mantle of the Archean Superior Province as imaged by garnet xenocryst geochemistry: Chemical Geology, v. 207, no. 3, p. 189-221.

Shen, W., Ritzwoller, M. H., and Schulte-Pelkum, V., 2013, Crustal and uppermost mantle structure in the central US encompassing the Midcontinent Rift: Journal of Geophysical Research: Solid Earth, v. 118, no. 8, p. 4325-4344.

Stein, C. A., Kley, J., Stein, S., Hindle, D., and Keller, G. R., 2015, North America's Midcontinent Rift: When rift met LIP: Geosphere, v. 11, no. 5, p. 1607-1616.

Thurner, S., Margolis, R., Levander, A., and Niu, F., 2015, PdS receiver function evidence for crustal scale thrusting, relic subduction, and mafic underplating in the Trans-Hudson Orogen and Yavapai province: Earth and Planetary Science Letters, v. 426, p. 13-22.

Thybo, H., and Artemieva, I. M., 2013, Moho and magmatic underplating in continental lithosphere: Tectonophysics, v. 609, p. 605-619.

Van Lankvelt, A., Schneider, D., Biczok, J., McFarlane, C., and Hattori, K., 2016, Decoding Zircon Geochronology of Igneous and Alteration Events Based on Chemical and Microstructural Features: a Study from the Western Superior Province, Canada: Journal of Petrology, p. egw041.

White, D., Musacchio, G., Helmstaedt, H., Harrap, R., Thurston, P., Van der Velden, A., and Hall, K., 2003, Images of a lower-crustal oceanic slab: Direct evidence for tectonic accretion in the Archean western Superior province: Geology, v. 31, no. 11, p. 997-1000.

Whitmeyer, S. J., and Karlstrom, K. E., 2007, Tectonic model for the Proterozoic growth of North America: Geosphere, v. 3, no. 4, p. 220-259.

Williams. M. L. and Hanmer, S., 2006, Structural and metamorphic processes in the lower crust: evidence from the East Athabasca mylonite triangle, Canada, a deep-crustal isobarically cooled terrane. In Brown, M., and Rushmer, T. (Eds.), Evolution and Differentiation of the Continental Crust, Cambridge University Press. p. 231-267.

Young, G.M., 2013, Precambrian supercontinents, glaciations, atmospheric oxygenation, metazoan evolution and an impact that may have changed the second half of Earth history: Geoscience Frontiers, v. 4, no. 3, p. 247-261.

Zhang, H., Lee, S., Wolin, E., Bollmann, T. A., Revenaugh, J., Wiens, D. A., Frederiksen, A. W., Darbyshire, F. A., Aleqabi, G. I., and Wysession, M. E., 2016, Distinct crustal structure of the North American Midcontinent Rift from P wave receiver functions: Journal of Geophysical Research: Solid Earth, v. 121, no. 11, p. 8136-8153.